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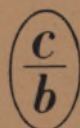
May, 1994

LITHOLOGY AND MINERAL RESOURCES

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A translation of *Litologiya i Poleznye Iskopaemye*

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GEOCHEMISTRY OF THE SEDIMENTARY PROCESS.

PART 2. TYPES OF SEDIMENTATION BASINS AND SOURCES OF ALIMENTATION AS FACTORS IN THE DIFFERENTIATION OF SUBSTANCES

V. N. Kholodov

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In the article, we discussed the relationship between phase differentiation of substances, on the one hand, and ultimate basins (lakes, seas, oceans) and composition of rocks composing catchbasins, on the other. It is shown that preferential weathering of basalts, granitoids, basitic-ultramafitic, or sedimentary rocks during the Earth's history resulted in the formation of different series of ore deposits and sometimes promoted the creation of global ore epochs. It is suggested that the rate of phase differentiation of substances during the sedimentary process increased from the Archean to the Quaternary.

In a previous article [112], it was shown that sedimentary differentiation of material in the zone of sedimentation is an extremely complex process in which mechanical, physical-chemical, and biogenic processes are interwoven. The quantitatively predominant mechanical redistribution of chemical elements forms the Clarke background against which ore concentrations are created via chemogenic or biogenic pathways. Their phase transformations, which are typically carried out under the influence of the active hydrodynamic environments of ultimate basins, play an extensive role in the precipitation, dissemination, and concentration of chemical elements.

Physical-geographic features of the environment (climate, relief) and sources of material (volcanism, distributive provinces) are very important in the sedimentary differentiation of chemical elements. As demonstrated by Strakhov [82], the sedimentation associated with each type of process (glacial, humid, arid, and volcanosedimentary) is characterized by its own order of Clarke or ore differentiation: a single scheme describing the separation of substances everywhere on the planet is a scientific abstraction.

In recent years, it has become clear that sedimentary differentiation of material largely reflects the morphometric types of ultimate basins, the chemistry and metallogeny of distributive provinces, and the secondary processes which take place during the later stages of sedimentary rock formation. The influence of these factors on the progress of sedimentary differentiation will be discussed in our forthcoming reports.

Types of Sedimentation Basins and Their Influence on the Completeness of Separation of Materials in the Zone of Sediment Formation

It is not very difficult to distinguish between lakes, inland and marginal seas, and oceans on the basis of morphometry and ratio between catchment and water areas. These three types of sedimentary basins substantially differ from one another [81].

Lakes are comparatively small water bodies. They reach 400,000 km² in area and only occasionally reach depths of 1 km. In terms of physical-geographic setting of the basin, mountain, plain, and coastal formations can be distinguished [73].

Most often, the lakes are located in areas of continental crust. Previous suggestions that lakes confined to continental rifts occur within areas of oceanic-type crust have not been confirmed by recent data [61]. The hydrochemical type of water in a lake reflects the features of climatic zoning to a significant degree. Occasionally, it can be linked to an abyssal flow of thermal water. In the work of Alabyshv [3], Strakhov [81], Kuznetsov [43], et al., it was shown that under conditions of hot, dry climate, lakes with highly mineralized waters of the sulfate, sodic, or chloride type are widely

occurring; whereas, in humid zones, analogous basins are filled with slightly mineralized waters containing elevated quantities of iron, phosphorus, SiO_2 , C_{org} , and other biogenic components. It is curious that diverse evaporites, magnesite, dolomite, anhydrite, gypsum, halite, etc., predominate among the bottom sediments in the first case; whereas sapropels, diatomites, and ferromanganese crusts and concretions predominate in the second case.

Sedimentation on the floors of lacustrine basins proceeds at a maximum rate; according to the data of Shostakovich [118], the average sedimentation rate reaches 9 mm/yr or 9 m/1000 yr. In terms of petrographic composition, lake sediments are fairly closely related to the rocks composing the catchment area. As shown by Nalivkin [56], the evolution of carbonate sediments, moraines, or effusive complexes in catchbasins is very sensitive to the overall composition of the lacustrine muds.

In size, seas differ very significantly from lakes. The maximum water area of the largest sea, the Mediterranean, is 2,856,000 km²; the average estimate of depth is 2-3 km.

In terms of position relative to ocean and continental block, we can distinguish between inland and marginal seas; in terms of morphometry, we can distinguish between flat-bottomed and trough seas [56, 81, etc.].

Seas occupy a fairly vague position with respect to predominant type of crustal structure: a number of seas of the Baltic or White type are located on continental crust. Other seas, such as the Red, Black, Caspian, and Japan seas, are located partially on continental crust and partially on oceanic crust.

The large water areas and substantial catchbasins of seas are typically subject to complex climatic effects. As in the case with the Black Sea [79], it is therefore frequently difficult to definitively assign an entire water mass to formations of a single climatic series. Moreover, as Nedumov [59, 60] has recently demonstrated, distribution of chemical elements to different parts of one and the same water body can occur under the influence of an arid (variegated distribution) or humid (smooth distribution) climate.

In marine water bodies, clear indications of climatic action are only evident in coastal zones; in pelagic areas of seas, indications of climatic zoning are weak and the overall pattern of sedimentation is smoothed out.

Typically, rates of sediment accumulation in seas are somewhat lower than those of lakes. In a monograph, Lisitsyn [47] showed that rates of sedimentation in seas vary from 0.01 to 0.4 mm/yr and average 4 m/1000 yr.

Only the coastal and shelf areas of a sea are maximally influenced by the petrographic composition of the catchbasin. In the deeper zone of a sea, the contours of mineralogical provinces are typically blurred [5, 6, 30, 96].

Oceans are the largest ultimate basins. The areas covered by water in these gigantic basins vary from 70 to 160 million km² and the average depth is 4-5 km.

Oceanic crust is most typical of ocean basins. The exposure of such crust on the floors of such basins at median ridges has a substantial influence on oceanic sedimentation and even on the overall mass balance in such basins. Typically, products of hydrothermal-eruptive activity (metalliferous sediments, "black smokers," and other phenomena complicating the development of the normal sedimentation process) are widely occurring here [11, 51, 83, 122, 131, 132]. At some sites in oceans, endogenous sources of material can therefore be correlated fairly well with exogeneous, incoming material from the continental block.

The influence of climate on oceanic bodies of water is even weaker than the influence on seas. However, certain features of sediments in abyssal zones reflect the climatic zoning of the planet [48]. The most important behavioral patterns of rock-forming components are very closely linked to the hydrodynamics of the water and are governed by acclimatic factors. Only within a narrow strip of coast does the oceanic sedimentary process clearly react to changes in average temperature and quantity of sediments. The influence of climate decreases in the direction of the pelagic zone and eventually becomes null [83-86].

The huge dimensions of oceans in comparison to seas markedly affect rates of sedimentation. It is only at discontinuities of the submarine shelf in the nearshore areas of these basins ("levels and belts of avalanche sedimentation" in the terminology of Lisitsyn [50]) that rates of sedimentation reach substantial values; over extensive pelagic regions, they exhibit a typical range of 2-3 mm/1000 yr [47].

This reduction of the sedimentary process in oceans is undoubtedly a consequence of the huge water-covered areas of these basins and their colossal depths.

It is quite apparent that the different morphometries, different ratios of areas of catchbasins to areas of water bodies, and also the different intensities of the sedimentary process in lakes, seas, and oceans generate very different types of sedimentary differentiation of substances.

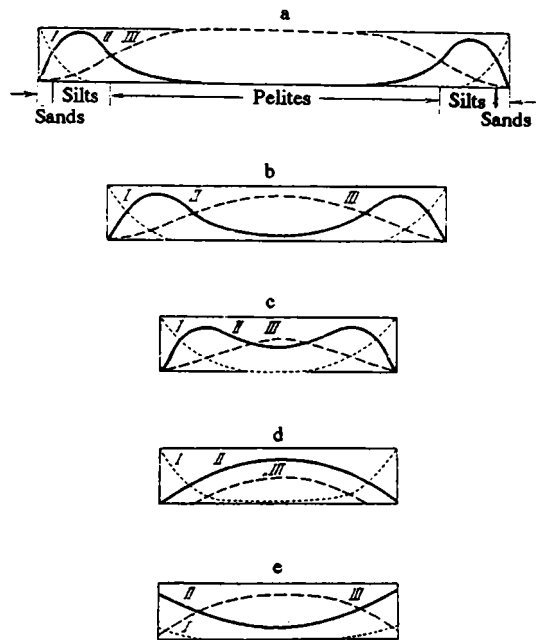


Fig. 1. Differences between mechanical differentiation of material in lakes, seas, and oceans [80]: a) ocean basins; b-e) intracontinental basins: b) 1st type (the Black and Aral seas), c) 2nd type (based on the data of M. V. Klenova), d) 3rd type (southern part of Lake Baikal), e) 4th type (Lake Balkhash). I-III) Fractions: I) sand, II) silt, III) pelite-clay.

Actually, the granulometric sorting of sediments in the "lake-ocean" series increases substantially with increasing size of the water body. Figure 1 shows the distributions of clastic particles in sand (>0.1 mm), silt ($0.1-0.01$ mm), and pelite (<0.01 mm) fractions in present-day water bodies of various types.

Note that two types of distributions of granulometric fractions are found lacustrine sediments. In the first case, in Lake Balkhash for example, a sand fraction is almost completely absent or accumulates sporadically along the shore; the maximum silt fraction is located along the shore and rapidly decreases toward the pelagic regions. The clay fraction is distributed in the reverse manner, however. Be that as it may, mixed clayey-silty sediments accumulate everywhere on a lake floor. In the second case, typical of Lake Baikal, predominantly silt fractions accumulate in the deeper parts of the water body beyond the zone of sands along the shoreline, although the quantity of pelitic material increases in the same direction. A predominance of silty-clayey muds is found over a substantial area of the bottom.

Mechanical differentiation in marine water bodies has a substantially different appearance. As a rule, zones of sandy, silty, and clayey sediments succeed one another from the shore to the depths of the basin; in such large water bodies as the Black Sea, separation of material into fractions is especially pronounced.

In oceans, separation of sediments material according to granulometric composition reveals yet another variant. As in seas, zones of sandstone, siltstone, and clay are clearly distinguishable, but regions with coarse fractions (sand and silt) are markedly restricted in the basin and are shifted toward the shore. Pelite can become predominant. It is precisely the clay fractions which have the greatest distribution in the central areas of oceans and which are typomorphic sediments of these huge water bodies.

It is easy to see that mechanical separation of sedimentary material is slightest in lacustrine sedimentation. The differentiation and integration of material in these basins is intricately interwoven, which results in the formation of granulometrically complex sediments. Granulometric mixtures often form here as are poorly sorted, thin-layered muds.

Mechanical differentiation of sedimentary material is much more clearly expressed in the seas, where well sorted sandy, silty, and clayey muds are often formed under the influence of active hydrodynamics.

TABLE 1. Distribution of Chemical Elements with Respect to Granulometric Fractions of Recent Sediments in Inland and Marginal Seas

Sea	Sand-silt fraction	Clay fraction	Source in literature
Black	Ti, Zr, Ge, Cr, V	Fe, Mn, Ni, Cu, Co, Mo, W, As, U, P, C _{org} , Au, Sr	[90]
Okhotsk	Ti, Fe, V, Cr	W, Cu, Ni, Mn, Mo, Co	[88]
Baltic	SiO ₂ bulk, Sn, Zr, Ti (?)	Fe, Ti, P, Zn, Cu, Cr, Ni, Co, W, Ba, V, Mn, Mo, Cd, C _{org}	[32]
White	Zr, Ti, Cr	Fe, V, P, Ni, Y, Co, Mn, C _{org}	[57]
Mediterranean	SiO ₂ , Sn, Zr, Cr, Ti, in part P	Fe, V, Mn, Ni, Co, Be, B, Zn, Cu, Au, Ag, C _{org} , etc.	[34]

Finally, in ocean basins (in which uncompensated sedimentation is especially characteristic), mechanical differentiation is most completely realized and, as well demonstrated by Strakhov [83, 87], frequently results in fractionation of the most pelitic sediments. Clayey sediments or sediments of miscellaneous granulometric composition form here and in the pelagic region.

It is generally accepted that clearly defined and invariant relationships hold between the granulometry, mineral composition, and chemical composition of muds [58, 65, 80, 113]. It can therefore be assumed that chemical differentiation of matter in lakes, seas, and oceans is also realized to different degrees of completeness.

Actually, investigations of distribution patterns of chemical elements in the Baikal and Balkhash lakes and in the Aral and Caspian seas carried out by Knyazeva [38], Sapozhnikov [72], Brodskaya [16, 17], Strakhov [80], Turovskiy et al [97], Baturin [7], Khrustalev et al [117], and Kholodov et al [114-116] allow us to assert that such patterns are usually very intricate. The frequent predominance of such diluting components as carbonates, silica, evaporite salts, and ferromanganese minerals, the formation of which, in lakes, depends upon the physical-geographic setting in which they are located, as well as variations in the composition of rocks of catchbasins, create a very broad range of concentrations of chemical elements. As a result, ratios of percent contents of chemical elements are peculiar to the muds of a given lake. For example, the content of iron in the sandy sediments of Lake Aral averages 0.88%; whereas, the content of iron in analogous muds of Lake Baikal reaches 3.09%.

A substantial portion of the chemical elements in lacustrine sediments are concentrated as a result of the smaller dimensions of the predominant fraction; another group of elements behaves independently of granulometry or even accumulates in sandy varieties. However, the compositions of these three groups of chemical elements typically vary from one water body to another. The behaviors of such chemical elements as Ti, Ba, V, Pb, and Ga vary especially frequently.

Although the behavior of the majority of chemical elements in lakes can be accounted for by classifying them into groups of elements exhibiting greater or lesser mobility, it is undoubtedly the case that a clear pattern of phase differentiation of matter is lacking in these water bodies.

The differentiation of chemical elements in inland and marginal seas is substantially more complete. In these water bodies, it is easy to identify a peripheral, nearshore area, in which sediments predominantly composed of sandy-silty fractions show maximum occurrence (Table 1). In this hydrodynamically active portion of the water body, we find accumulations of those elements — hydrolyzates which bind with the most stable host minerals. The actual set of chemical elements concentrated in the nearshore zone frequently depends upon the petrographic composition of the catchbasin. For example, the contents of Ti, Fe, and V in the nearshore parts of the Okhotsk Sea are generally determined by the widespread occurrence of basaltoids containing vanadium-bearing titanomagnetites in the nearshore areas of islands of the Kuril range. On the other hand, accumulation of Zr and Sn in the coarse-grained sediments of the Baltic Sea reflects the widespread distribution of acidic intrusions containing elevated quantities of accessory cassiterite and zircon throughout Fennoscandia.

However, the most characteristic feature of phase differentiation in seas is the presence of a very deep water region within which pelitic sediments substantially enriched in organic matter are typically deposited. A fairly large set of chemical elements, among which Fe, Mn, Co, Ni, Mo, U, and P are perhaps most typical, accumulate with them.

Separation of substances is even more marked in oceans. As shown in [27, 31, 33, 34, 52, 53, 83, 87, 91, etc.], the patterns of concentration of chemical elements are substantially different within ocean basins. Here, we can distinguish three groups of chemical elements which exhibit very different behavior in a lithological-facial profile.

The first group includes elements-hydrolyzates closely bound with stable host minerals. In oceans, as distinct from seas, they do not form a continuous zone within which they accumulate everywhere. In the nearshore zone of ocean basins, concentrations of Fe, Ti, V, Cr, TR, Zr, Hf, Ge, Pt, Au, etc. are sporadically created. Often, such concentrations are genetically linked, via corresponding distributive provinces, to the shore.

The accumulation of these elements in nearshore and shelf facies of present-day oceans is genetically linked to processes of placer formation. The latter were investigated in detail in the work of Petelin [64], Velichko and Korbut [21], Aksenov [2], Moore [129], Kronen [42], Aynemer and Konshin [1], and others.

It has been shown that two among the six different mineral associations of oceanic placers are the most abundant: 1) ilmenite, rutile, zircon, magnetite, garnet, monazite, occasionally sillimanite, cassiterite, staurolite, spinel, and disthene; 2) titanomagnetite, magnetite, pyroxene, occasionally sphene and apatite [1]. In the Pacific Ocean, placers are widely occurring along the eastern coast of Australia, along the shores of New Zealand, in the shoals of islands in the Zondsk Archipelago, in Indochina, along the coasts of the Japan Islands and Kamchatka Peninsula, along the southern shores of Alaska, and along the coasts of California and Oregon. In the Indian Ocean, regions with accumulations of minerals in placers are found along the coasts of the Republic of South Africa, Mozambique, Tanzania, and along the shores of Madagascar, the west and east coasts of India, in Indonesia, and in Western Australia. Finally, commercial deposits of placer minerals in the Atlantic have been discovered along the coast of Florida (USA), on the shelf of Brazil and Uruguay, and in the nearshore areas of Senegal, Sierra-Leone, and Morocco.

Thus, elements of the hydrolyzate group generally accumulate in a narrow and discontinuous circumcontinental belt girding a continent and along the coasts of islands under the influence of the sorting action of waves.

The second group are biogenic chemical elements, including C_{org} , $CaCO_3$, SiO_2 , N, S, Se, Br, I, etc. They accumulate in a more seaward direction, occasionally close to shores, and clearly gravitate toward shallow-water sediments [91], sometimes forming sublatitudinal belts crossing regions of various depths, as though accumulating over the entire facial plan of sediments [45, 47, 83].

Investigation of the patterns of behavior of these elements carried out with the aid of the percent method of absolute weights by Strakhov [83] indicate that the main source of the chemical elements in this group is plankton. This conclusion was later confirmed by the work of Bogdanov, Gurvich, and Lisitsyn [12, 13].

Since the biomasses of phyto- and zoo-plankton are higher in the peripheral areas of oceans and lower in zones of halistasis [15, 39], the overall plan of elements accumulation on the ocean floor conforms to this pattern to a substantial extent.

Actually, as was well demonstrated by Ramankevich [71], maximum contents of C_{org} in bottom sediments of oceans distinctly concentrate toward shelf shallow-water deposits and form a narrow, discontinuous circumcontinental zone extending along the coasts of the continents. Along with this, but to a lesser degree, the broad zoning generally characteristic of plankton and suspended organic matter is reflected in the distribution of organic carbon. The lower boundary of the distribution of organics in sediments is determined by the rates of oxidation and mineralization of buried planktogenic organisms, dissolved OM in the sea water, and also the length of the trophic chain and intensity of zooplankton transport [24, 71, 76].

Carbonate accumulations are substantially more widely distributed than organic carbon. Since their lower boundary of accumulation is limited by the carbonate compensation depth, which varies from 3650 to 6900 m in various parts of the ocean [46, 47, 83], the regions of their accumulation often encompass not only areas of the shelf and continental slope, but also zones of submarine rises.

Although the absolute weights of carbonate materials also tend to increase toward shallow water and decrease toward pelagic regions [47], the overall distribution pattern of carbonates in oceanic sediments is substantially more complex than the accumulation pattern of carbon.

The behavioral pattern of SiO_2 is still more intricate. Solution processes have substantially less effect on its precipitation, as a result of which many siliceous shells of radiolarians and diatoms reach the deepest areas of the bottom. Their accumulation correlates well with climatic zonation [45] or hydrodynamics [83].

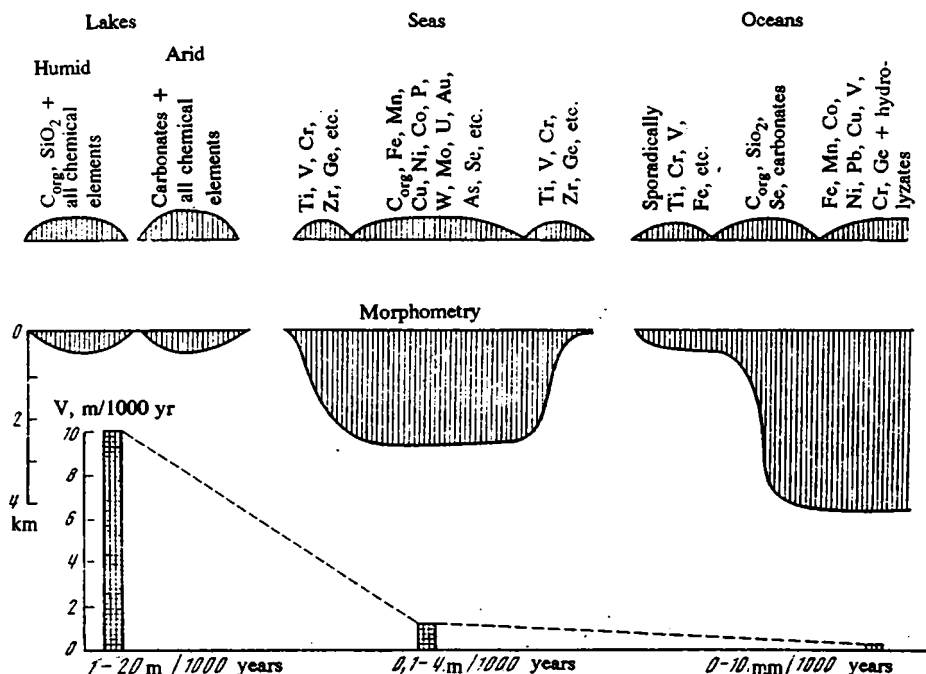


Fig. 2. Dependence of chemical and biogenic differentiation upon sedimentation rate and morphometry of water bodies.

Strakhov [83] discovered a characteristic feature of the behavior of biogenic chemical elements in the oceans. He found that their distribution correlates slightly with the granulometry of the sediments. In this respect, oceanic sedimentation exhibits a cardinal difference from marine sedimentation, since, in marine sediments, the content of chemical elements in the biogenic group is closely linked to the grain sizes of the terrigenous particles in the sediment.

The mechanism controlling the behavior of biogenic elements in the ocean, in all its steps, is the life activity of phyto- and zooplankton. Initial concentration of chemical elements takes place from solutions and suspensions. The most important distribution patterns of these components on the bottom are, to a significant degree, results of the distribution of plankton biomass in the topmost layer of ocean water. The accumulation of plankton in sediments depends upon the stability of biogenic formations and upon their participation in trophic transformations [9, 12, 14, 15, 23, 49, 83, etc.].

Elements associated with iron and manganese belong to the third group. According to data gathered by Strakhov [83], they form four subgroups: essentially ferric (Fe, Ge, V, Cr), hydrolyzate (Al, Ti, Zr, Ga, Ta, Nb), calcophile (Pb, Zn, As, Cu), and manganic (Mn, Co, Ni, Mo). These include primarily hydrolyzates, either incorporated into the compositions of clay minerals or precipitated in hydrogenic forms.

The bulk of the elements in the iron-manganese group accumulate in the deepest parts of oceans and near median ridges. They are spatially and genetically associated with fields of metalliferous sediments and concretions consisting of ferromanganese hydroxides, but, despite the different sources of the metals (demonstrated by Strakhov, Volkov, and Lisitsyn [91]), the bulk of the elements in this group are distributed along the bottom in complete accordance with the granulometry of the terrigenous portion of the sediments. In addition, analysis of the behavior of absolute weights of the chemical elements indicates that maxima of elements are offset relative to one another from the coast to the pelagic region and form predictable zones of accumulation of first ferric, then hydrolyzate, calcophile, and manganic groups [83]. This especially high degree of differentiation of matter is beyond what can be reached under conditions found in seas.

The mechanism of distribution of elements in the iron-manganese group is closely linked to phase transformations and hydrodynamic fractionation of solid particles. According to Strakhov [83], dissolved elements of the iron-manganese group brought to the ocean from the land rapidly coagulate and the condensations behave as a typical mechanical suspension. Evidently, the fate of elements entering the water via the hydrothermal-sedimentary route is analogous. Plankton complicate this process by partially absorbing, assimilating, and again dissolving these components in sea water. Repeated passages from suspension to solution and back, taking place against a background of hydrodynamic sorting of material, eventually results in very thorough separation of components and creation of fairly well defined chemical-granulometric series; the colossal scales of ocean basins and the active hydrodynamics of the water mass facilitate this process.

It is generally accepted that all of the behavioral patterns of chemical elements discussed above characterize the clarke process taking place in lakes, seas, and oceans; the conditions of formation and distribution patterns of ore accumulations are distinguished by greater complexity; they have been discussed in [112] in general terms.

By comparing processes of present-day clarke differentiation taking place in lakes, seas, and oceans, it is easy to see that these processes clearly become more complicated as we move from small water bodies to large ones (Fig. 2).

In lakes, we find weak separation of components which is frequently distorted by climatic pressures and dilution of allochthonous material by autochthonous materials such as salts, evaporites, sapropel, diatomites, etc. This can completely hide the overall geochemical patterns associated with the process.

In seas, phase differentiation becomes the main factor in sedimentation. Mechanical fractionation, which, in essence, obliterates indications of the source of formation of the solid phase, combines the supply of suspensions, biogenic sedimentation, and physical-chemical processes associated with the formation of the solid phase and balances the distribution of terrigenous, biogenic, and chemogenic components.

In oceans, separation of genetically diverse particles reaches its highest phase. Despite a greater influence of volcanosedimentary processes, rates of sedimentation in these water bodies are generally less than rates in seas and biogenic factors are most important. The biogenic differentiation in oceans is the most complete type, and, in the shallow zone, competes fairly successfully with processes of mechanical fractionation. In the deeper parts of oceans, muds devoid of organic matter and biogenic elements but substantially enriched in elements of the iron-manganese group form under the influence of phase differentiation. These muds serve as indicators of recent oceanic sedimentation [92].

Thus, it is apparent that a low rate of sedimentation and a large ultimate basin allow maximally complete separation of material and promote very distinct phase differentiation.

Recently [92, 93, etc.], attempts have been made to find major trends in the evolution of sedimentation basins and the volume of the hydrosphere over the Earth's history. It has been found that the structure of the lithosphere through the admittedly incomplete geological record leads to the following conclusions:

1) there is an absence of clear indicators typical of present-day oceanic sedimentation: red deep-water clays, foraminiferan and coccolithic-foraminiferan clayey-carbonate clays as well as radiolarian, diatom, and ethmodiscus siliceous and siliceous-clayey deposits formed in environments of extremely slow sedimentation and containing an association of ferromanganese concretions, zeolites, celestobarite, and Fe-smectite in beds of sedimentary and volcanosedimentary rocks of the early Mesozoic;

2) there is a predominance of relatively shallow-water and terrestrial deposits (paleosoils, zones of weathering, evaporite lenses, stromatoliths, and reefogenic formations with dissication cracks, alluvial-lacustrine cross-bedding, and other shallow-water-litoral structures) in Precambrian and Paleozoic sections;

3) there is an increase in the contrast of the surface relief from the Precambrian to the Quaternary; this is reflected in an increased rate of sediment transformation in monotypical sea beds, in a reduction in thalassocratic areas within the Earth's continental block, in an increase in the abundance of rudaceous rocks (psephites), and a number of other lithological-facial indices of the lithosphere;

4) the pattern of decelerating and accelerating growth of the hydrosphere from the Precambrian to the Quaternary period was probably due to an imbalance between the supply of water from the degassing mantle and consumption of water during the serpentinization of basalt, granitization, and hydration of sedimentary strata. In the final analysis, the present mass of the hydrosphere ($1.37 \cdot 10^{24}$ g) was formed only during the Jurassic or Cretaceous.

Taken as a whole, the data present allow us to conclude that the ultimate basins on the Earth's surface underwent a complex evolution from shallow-water lacustrinelike water bodies of the Precambrian through Paleozoic seas to Mesozoic-Cenozoic oceans. In essence, present-day oceanic sedimentation is probably the highest state in the evolution of ultimate basins; its beginning can be dated as early Mesozoic.

If all this is true, clarke phase differentiation of material should have undergone substantial alterations over geological time. In ancient Precambrian lakelike basins, separation of material mobilized on continents took place to an especially low degree. At this time, thin-layered multicomponent sediments of the jaspilite, phthanite, and itabirite types became widely distributed. These mixtures of SiO_2 , Fe, Mn, and organic matter, often adjacent to muds, were facially and geochemically similar to recent shallow-water and lacustrine deposits. Later, during the Proterozoic and Paleozoic, phase differentiation increased and resulted in the formation of well specialized sediments of the marine type. Finally, the most

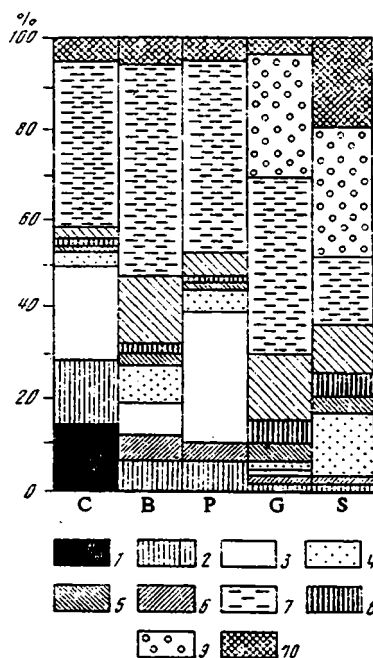


Fig. 3. Chemical compositions of various petrographic types of rocks and meteorites: 1) metallic iron; 2) FeO; 3) MgO; 4) CaO; 5) Na₂O; 6) Fe₂O₃; 7) SiO₂; 8) K₂O; 9) silica occurring in the form of quartz; 10) accessory elements. C) Stone meteorites (chondrites); B) basalts; P) peridotites; G) granites; S) sedimentary rocks.

complete manifestation of phase differentiation began to take place during the Mesozoic–Cenozoic, when oceans reaches their widest development on the Earth's surface, the hydrosphere assumed its present appearance, and separation of chemical elements reached its zenith [125].

Geochemical Types of Distributive Provinces and Phase Differentiation of Material

Associations of chemical elements which play roles in the phase differentiation of material are substantially restricted by the composition of parent rocks, which are subject to processes of weathering and erosion over the area of a catchbasin.

In turn, the proportions of various types of disintegrating parent rocks substantially depend upon the stages of development of each of the fold systems present. Actually, the initial stages in the emplacement of orogens typically result in erosion involving those young sedimentary and volcanosedimentary strata which compose the peripheral parts of a mountain structure. As erosion and peneplanation in the sedimentary cycle deepen, old, metamorphosed, and, often, igneous and vulcanogenic rocks composing the central parts of mountain systems become progressively more involved; they create a pattern of appreciable and always increasing influence of juvenile material on the course of the sedimentary process within a given region.

Despite the fact that massifs of igneous rocks in the cores of mountain structures typically outcrop over relatively small areas, the geochemical and metallogenic significance of such rocks is very great. For example, the narrow chain of intrusions of gabbro, pyroxenite, dunite, peridotite, and other basites and ultramafites outcropping in the present-day section affected by erosion in the Hercynian of the Urals contains huge amounts of chromium, iron, titanium, vanadium, copper, phosphorus, and platinoids. It has actively affected the geochemistry of the sedimentation in this province beginning with the Permian. This influence has continued into the present [98, 105].

TABLE 2. Absolute Weights of Chemical Elements in Rocks of Various Types, kg/ton [22]

Element	Basic and ultrabasic rocks (dunites, peridotites, basalts, gabbro)	Acidic rocks (granites, liparites, etc.)	Sedimentary rocks (clays, schists, sandstones, etc.)
Mn	1,3-2,2	0,6	0,6
P	1,2-1,4	0,7	0,7
Cr	0,3-2,0	0,025	0,16
Ti	3,0-9,0	2,3	4,5
Ni	0,16-1,2	0,008	0,095
Co	0,04-0,2	0,005	0,023

The facts presented above suggest that it would be interesting to compare the mineral-geochemical composition of igneous, metamorphic, and sedimentary rocks. Figure 3, which is taken from the work of Kropotkin [41] and slightly modified on the basis of other published data, provides a comparisons of average chemical compositions of chondrites (C), basalts (B), peridotites (P), granites (G), and sedimentary rocks (S).

Analysis of the graph indicates that the greatest differences are observed in the average contents of iron, magnesium, and silica. Actually, the quantity of divalent and native iron, reaching a total of almost 30% in chondrites, basalts, and peridotites, shows a drop by a factor of almost two, with hydrous ferric oxide appearing instead of native iron. A further decrease in total content of mixed iron oxides occurs in granite and sedimentary rock; here, it again decreases by almost 50% relative to basalts and peridotites.

The content of MgO typically varies from 7 to 30% in stony meteorites, peridotites, and basalts.

Finally, let us turn our attention to the behavior of silica. The highest quantities of silicate, i. e., the SiO₂ showing the greatest tendency to migrate, is found in basalts, chondrite meteorites, and peridotites. The average content of active silica in these rocks is sharply reduced as a result of the wide distribution of weakly eroded, very stable quartz in granites and, especially, in sedimentary rocks.

At the same time, it should be emphasized that the weight of "accessory" elements, among which H₂O, CO₂, C_{org}, etc. play important roles, increases in sedimentary rock; they reflect the participation of the hydrosphere and biosphere in the formation of the sediments.

Thus, basic and ultrabasic igneous rocks are the main bearers of iron, silica, and magnesium. They primarily supply these components to neighboring areas of sedimentation by involvement in processes of weathering and hypogene decomposition.

Many ore elements behave in similar fashion, especially those which are usually paragenetically associated with iron; these include Mn, P, Cr, Ti, Ni, Co, V, and others. The majority of these ore elements typically show sharply elevated concentrations in basic and ultrabasic igneous rocks. This is well illustrated in Table 2, where quantities of these elements (in kg) found in 1 ton of parent rock are calculated from crustal abundances obtained from a report by Vinogradov [22]. Apparently, Mn, P, Cr, and Ti, are 2-3 times more concentrated and Ni and Co are 4-5 times more concentrated in basites and ultrabasites than in granitoids and sedimentary rocks.

While basites and ultrabasites are the main carriers of metals in the series Mn, P, Cr, Ti, and Ni, the geochemical distribution of phosphorous is more complex; its average content in basites typically ranges from 0.17 to 0.47 wt. % [22, 99, 132]; in alkaline rocks, the average content of phosphorous shows a wider range of variation: from 0.03 to 0.64% [28]; it reaches 5-6% in carbonatites spatially associated with alkaline rocks [67, 77].

In addition to showing a pattern of high clarkes, basites and ultrabasites are spatially associated with a group of so-called magmatic deposits; these include accumulations of chromite, vanadium-bearing titanomagnetite, platinoids, and sulfides of copper, cobalt, and apatite [20, 62, 77].

The ores in deposits of this type are usually confined to the body of the intrusion itself, where regular beds and lenticular areas form and vein deposits and irregularly shaped deposits are less common. The maximum contents of metals in such deposits reach very significant levels. For example, the content of Cr₂O₃ in chromite beds and zones can reach 45-50%; the content of iron in titanomagnetite ores can reach 60%; that of titanium dioxide can reach 20%; that of vanadium can reach 1.5-2% [107]. In the platinum-bearing zone of Merensky Reef (the Bushveld complex, Republic of South Africa), variations in contents of precious metals equal (in mg/ton): iridium, 0.2-0.6; ruthenium, 3-12; platinum, 30-50;

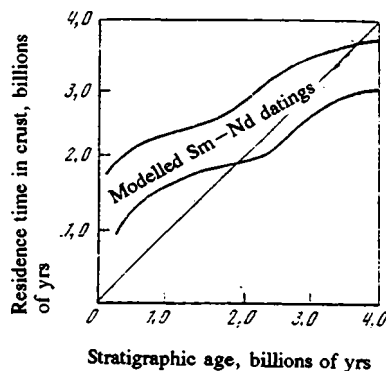


Fig. 4. Relationship between modelled Sm – Nd datings of clay shales and their stratigraphic ages [100].

gold, 10-40 [133]. The quantity of nickel in aveolar ores of the Nickel Mountain deposit reaches 1.0-1.5% [36]. Finally, the complex apatite-magnetite ores of the Kruchinin deposit (the Angashansk pluton, Russia) contain 70-90% ore minerals and up to 30% apatite [26]. It is undoubtedly true that this entire mass of ore matter, after being subjected to processes of weathering and sediment formation, created very special conditions for sedimentary differentiation. This began with the mobilization of very high ore concentrations which then and there ensured the formation of sedimentary deposits of various metals [106, 107, 110].

The main types of igneous rocks in the centers of disintegrating mountain structures did not remain the same throughout Precambrian and Phanerozoic geological history; the different types succeeded one another in complete accord with the laws governing the evolution of magmatism, which were investigated in a series of earlier reports [10, 106, 108, 110].

Although the oldest distributive provinces in Earth's history are not preserved, there is indirect evidence of an initial "lunar" stage; basalts completely predominated in areas with catchbasins during this stage and a powerful impression was made upon the overall geochemical character of sedimentation processes.

Recently, the existence of an ancient epoch during which mantle material exerted a dominant influence on Archean and Proterozoic sedimentation has been directly confirmed through an analysis of samarium-neodymium datings of clay shales carried out by O'Nions et al. [130], Allegre and Rousseau [121], and Goldstein [126], and reviewed in a book by Faure [100]. The results of these investigation are shown in Fig. 4.

The diagram is based upon numerous determinations of $^{147}\text{Sm}/^{144}\text{Nd}$ ratios carried out for noncoeval clay shales in different regions of the world (Great Britain, The Republic of South Africa, etc.). Clay shales were investigated on the basis of the proven fact that the ratio Sm/Nd in such rocks is unaltered over the course of weathering of the original material of the parent rocks and over the course of transport, deposition, and diagenetic alterations of the sediments. It always reflects the ratio Sm/Nd in the parental igneous rocks.

The diagram shows that clay shales formed from 2 to 4 billion years ago were, for the most part, created directly from mantle material: beginning 2 billion years ago and continuing into the subsequent Phanerozoic, the crustal source had an ever increasing influence on Sm/Nd ratios modelled. This resulted in deviations from a direct proportion.

From Fig. 4, one can thus fairly accurately trace the ancient epoch of sedimentation when mantle basalts were the main source of sedimentary material on the planet. It can be assumed that their erosion initially resulted in the formation of greywacke sandstones, especially typical of ancient greenstone belts, and that later intensive weathering and decomposition in an oxygen-free atmosphere sharply enriched in carbon dioxide [25, 37, 102] caused massive efflux of SiO_2 and many ore components (iron, manganese, phosphorous, titanium, vanadium, etc.) from the weathering zones created and resulted in accumulation of these substances within areally extensive but shallow water bodies. The iron-ore belts of the Earth where there are jaspilites of the Verkhoe Lake type, clearly confined to a period 2.6-2.0 billion years ago, formed as a result.

We described structural features of deposits formed during the Lower Proterozoic epoch of ore formation in a series of previous reports [109-111].

Jaspilites are frequently referred to as "banded" iron ore and manganese-ore formations because of their characteristic structure: they typically consist of an alternation of layers of silica, occasionally carbonates (dolomites), clayey or carbonaceous material with flakes of ore minerals of iron, manganese, and phosphorous. Here, iron in jaspilite rocks is typically represented by magnetite, hematite, martite, siderite, ankerite, greenalite, stilpnomelane, chlorite, minnesotaite, or pyrite; manganese is represented by manganite, pyrolusite, birnessite, cryptomelane, rhodochrosite, and other minerals; phosphorous is represented by apatite and its finely dispersed varieties.

Micro- (0.2-2 mm), meso- (1 mm to several cm), and macro-layering (several meters) can be fairly clearly identified in the most typical sections of ore beds; the sequence of thin layers often forms regular cycles and is rhythmical. According to the data of Trendall [94], individual microlayers in the Hamersly iron-ore formation (Australia) can be traced for a distance of 296 km; in many other Proterozoic iron-ore basins of the world, intensive folding and fault tectonics prevents precise correlation of sections.

According to the ideas of Plaksenko [65], Mel'nik [55], Trendall [94], Fralik [124], Braterman and Cairns-Smith [123], etc., the rhythmical alternation of thin layers reflects the periodicity of supply of silica, carbonates, and ore components from the continent, and well as seasonal fluctuations in biogenic phenomena.

Jaspilite deposits are typically represented by a complex association of ore-bearing and ore-free formations; according to Formozova [101], primarily formations of taconitic, itabiritic, and Krivoy Rog types formed during the Lower Proterozoic epoch.

Taconitic formations are represented by repeated alternations of jaspilite, chert, graphite-pyrite shale, greywacke, and clay shales; carbonates are usually not characteristic. In addition to ferruginous quartzite (earlier), there are oolitic iron ores and many silicites with algal structure. The iron deposits of Lake Superior (USA) and also the Hamersly region (Australia) are typical examples of this type.

Iron-ore formations of the Krivoy Rog type are similar to taconitic formations. In addition to jaspilite, they contain various shales, phyllites, and conglomerates. Phosphate and siliceous-carbonaceous rocks also enter into the association. This group includes deposits of the Kursk magnetic anomaly and Krivoy Rog (the Ukraine).

Formations of the itabiritic type are special. In addition to ferruginous quartzite, they consistently contain iron-manganese and dolomite layers rich in iron, manganese, and silica; accumulations of riebeckite can often be found. Itabirites are typical of the southern continents; they occur in Brazil and the Republic of South Africa.

Numerous lithological indications of extremely shallow water are usually evident in jaspilites of iron-ore formations. Evidence of the formation of zones of weathering and widespread occurrence of alluvial quartz conglomerates are typical of the underlying strata; dissication cracks are typical of carbonate deposits. It is characteristic that zones of weathering and auriferous alluvial conglomerates are usually paragenetically associated with iron deposits in many countries of the world [40, 70].

Scour marks, intraformational unconformities, ripple marks, and cut structures are widely distributed in the jaspilite series themselves. Even pseudomorphs after gypsum and halite can be occasionally found. The presence of shallow-water oolitic structures and finds of numerous mineralized bacterial-algal remains and large stromatoliths are in fair accord with these features. Diverse concretionary structures and molds of dissication cracks are found in some granular and banded ferruginous quartzite [124, 128].

In general, it is evident that separation of substances in Lower Proterozoic time proceeded according to special patterns; it took place during an excess of ore components and associated components in catchbasins and also under conditions of an oxygen-free atmosphere. Here, the ore-forming process was confined to shallow waters but occupied extensive water areas [19, 44, 65, 66, 78, 94, 95]. Phase differentiation here clearly competed with integration, is incomplete in places, and very rarely reached the point of complete separation of ore accumulations from each other. Colloidal systems and biogenic processes played a major role in the separation of sedimentary material.

Tectonomagmatic events of the Upper Cambrian began to substantially alter the structure and composition of continental catchbasins. Already at the Archean and the Lower Proterozoic, the distributive provinces of certain areas of sedimentation turned out to be complex diverse granitoids, acidic and basic effusives, and different types of sedimentary strata.

In the sedimentation of this time, this was reflected in a widespread distribution of diverse arkosic sandstones and gritstones. Complexes of acidic magmatic and effusive rocks gradually formed the granitic-metamorphic basement of the protocontinents [10] and this, in turn, made possible the injection and emplacement of large basite-ultramafite intrusions, which superseded komatiites.

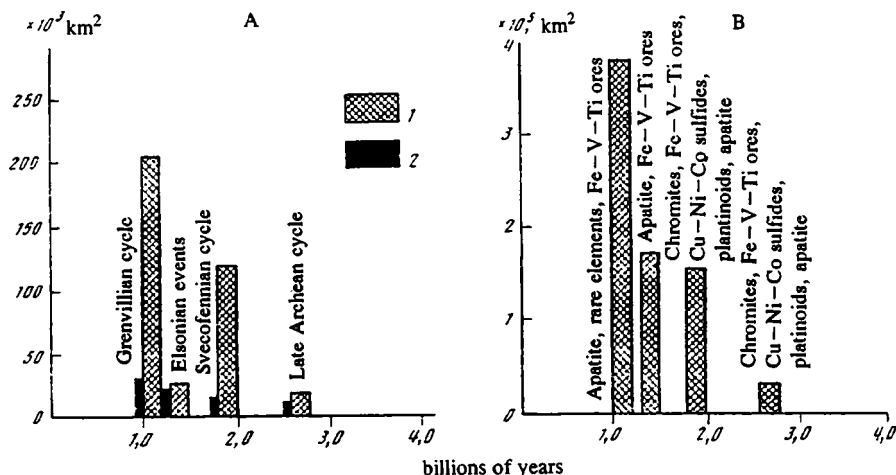


Fig. 5. Relative roles of basite-ultramafite and alkaline magmatism in distributive provinces of the Precambrian. A) Areas occupied by basite-ultramafite intrusions during the late Archean, Svecofennian, and Grenvillian cycles; B) growth rates of areas occupied by basite-ultramafite intrusions and their ore contents; 1-2) basite-ultrabasite and alkaline intrusions respectively.

During the Lower Proterozoic, there was intensive development of Archean folding (2.6-2.8 billion years) associated with the creation of granite, alaskite granite, granodiorite, and adamellite complexes. Potassic granitoids, rapakivi associated with ancient basite-ultramafite intrusions, are especially noteworthy [54, 110]. The late Archean phase of folding marked the beginning of the formation of the crust of the protocontinents [10]. Later, throughout the Upper Precambrian, tectonomagmatic events were repeated many times, but their scales, importance, and intensities varied substantially [74, 75].

The formation of archicontinents was completed with the Svecofennian cycle (1.8-2.0 billion years ago), at which time formation of granite-rapakivis, granites, plagiogranites, granodiorites, and basite-ultramafite intrusions was again taking place with intensity. In contrast to previous events, the subsequent Elsonian tectonomagmatic events (1.5-1.3 billion years ago) and, especially, the Grenvillian tectonomagmatic cycle (1.2-1 billion years ago) were not associated with the formation of new crust. They manifested within narrow belts and accelerated reworking of old crust [74, 134].

Since each tectonomagmatic cycle added a new portion of juvenile magmatic material to a region of folding and subsequent erosion, it would be especially interesting to characterize the petrographic compositions, areas, ages, and ore contents of basitic, ultramafic, and alkaline intrusions and to plot variations in these parameters over the course of the Proterozoic. For this purpose, we compiled published data characterizing 52 of the largest intrusions in the world. On the basis of this more or less precise geochronological data, we were able to establish correlations with Late Archean, Svecofennian, Elsonian, and Grenvillian tectonomagmatic events. The data are plotted in Fig. 5, where changes in total area of basite-ultramafite and alkaline intrusions over time are shown on the left and the dynamics of growth and evolution of the metallogeny of basite-ultramafite complexes over the Proterozoic are shown on the right.

Examination of the graphs leads to the following conclusions.

1. In terms of area occupied, the largest individual plutons of basite-ultramafite were emplaced during the Svecofennian and Grenville tectonomagmatic cycles (the Bushveld and Rizhsk intrusions, and also Shabogamo, Duluth, and Sen-Jin); maximum total areas occupied by basic and ultrabasic magmatites are characteristic of these events. The late Archean tectonomagmatic cycles, which, like the Elsonian, were distinguished by intrusive bodies of fairly small areal extent, stand in contrast to them.

2. Each Precambrian tectonomagmatic cycle or event ended with the formation of alkaline intrusions. Apatite-bearing carbonatites are frequently associated with the latter. In general, the abundance of alkaline rocks and carbonatites increased toward the middle of the Upper Proterozoic, reaching a maximum during the Grenville epoch.

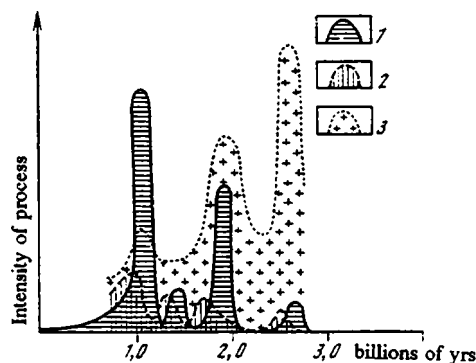


Fig. 6. Relationship between various types of magmatism and tectonomagmatic cycles in the Precambrian: 1-3) magmatism (1: basite-ultrabasite; 2: alkaline; 3: granitoid-acidic).

3. Areas occupied by basite-ultramafite intrusions reached maximum values in the interval from 2 to 1 billion years ago. Simultaneously with the increase in the sizes of basite-ultramafite intrusions, there was a decrease in the metallogenic diversity of such intrusions. For example, large ore deposits of chromite, platinoids, and Cu–Ni–Co sulfides are not found among Elsonian and Grenville basite-ultramafite intrusions, but, as in the preceding epochs, deposits of apatite and vanadium-bearing titanomagnetite ores continued to form [36].

4. Maximum apatite- and vanadium-bearing intrusive complexes emerged as distributive provinces at the end of the Precambrian.

The empirical patterns presented above complement Fig. 6, which is an attempt to evaluate the relative contributions of granitization processes as well as basite-ultramafite and alkaline magmatism over the entire course of the Upper Precambrian on the basis of purely semiquantitative data. The graph shows that the igneous component of Precambrian tectonomagmatic cycles underwent qualitative directional changes over time. The relative contribution of granite formation, which was very important during the later Archean, gradually decreased by the end of the Precambrian and was terminally reduced during the Grenvillian tectonomagmatic epoch. On the other hand, the rate of development of basitic magmatism, also including ultrabasic and alkaline rock complexes, steadily increased within the same trend, reaching its most complete manifestation during the Grenvillian diastrophism.

There is every reason to assume that the ratio of areas occupied by basite-ultramafite and alkaline igneous rocks, on the one hand, and granitoids, on the other, reached maximum values within the continent block of the Earth by the end of the Precambrian. The fact that the rate of breakdown of basite-ultramafite complexes in the zone of hypergenesis exceeded the rate of breakdown of granitoids by a factor of 5-6 also played a major role in the erosion of uplifted areas [35, 68, 69, 135]; moreover, according to experimental investigations by Pedro [63], the breakdown of basalts proceeds at a rate 11 times faster than the breakdown of granites. It is easy to conclude that material supplied from the weathering of basic-ultrabasic and alkaline igneous rocks should have dominated the composition of the distributive provinces.

The Vendian–Cambrian epoch of sedimentary ore formation, investigated in detail by Al'tgauzen [4], Gimmel'farb [29], Yanishin [119], Bushinskiy [18], Kholodov [103-105, 108, 110, 111], and Yanshin and Zharkov [120], was a reflection of these processes. During the Vendian–Cambrian, a typical association of ores and rocks was abundant in the very diverse marine paleobasins of the world. Bedded and nodular phosphorites, "black" coal-bearing shales, and phanites with concentrations of V, Cr, Co, Ni, Mo, U, Se, and other metals, deposits of iron and manganese ores, and also tillite and tillitelike rocks, dolomite, limestone, silica, and quartzite were frequently formed here. Among the deposits and ore shows of phosphorous, iron, and manganese, two somewhat different groups can be distinguished. In the first group, the ore component consists of segregated beds or lenses confined within strata of the enclosing rock (the phosphorites of Aksay, Kok-Jon, the phosphorite-bearing basins of Kara Tau, Kazakhstan, the manganese ores of the Durovsk deposit, Salair Ridge, etc.). In the second group, the ore component is finely interlayered with silica, dolomite, or limestone,

forming ore accumulations reminiscent of Proterozoic jaspilites (the iron ores of Dzhemym Tau, Kazakhstan; Dunderland, Norway; the Uds-Selemzhinsk interfluvium, Russia; phosphorites of Kok-su, Zhanatas, Berkut, the phosphorite-bearing basin of Kara Tau, Kazakhstan; the manganese ores of the Usinsk deposit, the Lesser Khingan, Russia, etc.).

In terms of "metallogenic composition," the Vendian–Cambrian epoch of ore genesis generally resembled the Lower Proterozoic. However, there were substantial differences: whereas iron and manganese were the major ore components during the Proterozoic, with phosphorous, vanadium, and rare metals accumulating relatively rarely, phosphorous and vanadium-bearing shales undoubtedly predominated over iron and manganese during the Vendian–Cambrian [110]. As we pointed out above, this inversion of metallogeny can be explained by assuming an evolution in the composition of distributive provinces and the atmosphere during the Archean and Proterozoic.

In general, it should be noted that the chemical composition of Lower Proterozoic ferruginous quartzites is especially complex. Also, the engineering properties of these quartzites leave much to be desired. First of all, consider the fact that the quantity of SiO_2 in these formations varies from 6 to 70%; the content of iron is just as variable, ranging from 11 to 70%. Since, in jaspilite, SiO_2 is represented by quartz, while iron is represented by magnetite, hematite, and a number of other metals, the main forms of sedimentation of these components in ultimate basins remain hypothetical; nevertheless, the majority of authors [8, 44, 65, 78, 127] believe that these minerals formed as a result of recrystallization of colloids which coagulated in Lower Proterozoic shallow seas.

Locally, especially in jaspilites of the Republic of South Africa, Brazil, and India, there are carbonate deposits and facially associated manganese ores among ferruginous-siliceous strata; the content of Mn, which usually does not exceed hundredths or tenths of a percent, increases to 50% in such places.

Less commonly, one finds high contents of phosphorous in jaspilite deposits. Such high contents are typical of sulfide facies and are present in the KMA, in the deposits of Lake Superior, and in India. It is undoubtedly true that phase differentiation of ore material during the Lower Proterozoic was especially weak and complicated by diverse integration processes.

The chemical composition of Vendian–Cambrian deposits appears to be substantially different. First, consider the fact that the content of SiO_2 is less in ores of this age; it rarely exceeds 40%, although there is also reason to believe that, for the most part, the SiO_2 here was initially colloidal sediment [109]. Against this general background, the numbers of separate deposits of phosphorite (10–15% P), manganese ores (15–40% Mn), and ore-ore body (30–40% Fe) are sharply higher. From the Lower Proterozoic to the beginning of the Phanerozoic, segregation of ore accumulations as well as rate of phase differentiation of material apparently increased.

The increase in separation of components during the sedimentary process beginning in the Phanerozoic was apparently a result of several factors. On the one hand, substantially larger, deep-water, troughlike seas replaced the extensive but very shallow waters of the Proterozoic; on the other hand, the role of repeated reworking and "maturation" of sedimentary material increased. Finally, it is apparent that the relative importance of juvenile siliceous acid and, consequently, the ability of such acid to "dilute" ore components in sediments underwent a relative decrease with increasing numbers of sedimentary cycles. All of these events promoted the formation of geochemically "pure" ore deposits.

The process of increasing ore differentiation continued into more recent epochs. It is generally accepted, for example, that ever more segregated and chemically purer deposits of iron, manganese, and phosphorous, constituting fairly homogeneous accumulations of ore components, formed in a number of regions during the Tertiary.

For example, the Oligocene manganese ores of Nikopol (the Ukraine), Chiature (Georgia), and Mangyshlak (Kazakhstan) contain only 3–20% SiO_2 , while special investigations into the forms of silica present indicate that the bulk of silica is incorporated into the compositions of clay minerals or forms a primarily terrigenous admixture [89]. The quantity of Mn in the ores of these deposits usually ranges from 10 to 50%. An almost complete absence of iron (1–2%) is very typical.

The composition of the ores of the Kimmeriya Kerchensk iron-ore basin (Crimea) is no less telling. These ores contain, in %: Fe 35–40; SiO_2 6–20; and Mn no less than 2. This is fairly clear from the example of the brown and tobacco-colored ores of El'tigen-Ortel'sk and Kyz-Aul'sk troughs.

The conclusion is obvious: as in the case of the Clarke process, the rate of ore phase differentiation of material on our planet has probably increased from the Archean to the Quaternary.

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COMPOSITION AND CONDITIONS OF FORMATION OF THE MANGANIFEROUS MEMBER OF THE POD"EMSK SUITE OF THE UPPER RIPHEAN (YENISEY RIDGE)

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UDC 550.4:551.72(551.53)

In the article, we show that an ore band at the base of a carbonate – volcanogenic – terrigenous formation accumulated in a basin with a rugged bottom as a result of coprecipitation of ore suspensions, organics, acidic pyroclastics, and detrital material. Manganese was reduced in muddy waters and precipitated in carbonate phases cementing and replacing pyro- and lithoclasts.

In deposits of the Precambrian Yenisey Ridge, several levels with manganese content have been discovered [2]. The Pod"emsk suite, present in the northern part of the Vorogovka trough, is the most important of these. This orogenic structure has complex borders determined by block movements along fissures. The bedding of the deposits of the late Riphean is gently folded. The Porokinsk syncline, associated with smaller folds of northwestern trend, is the largest structure of its kind in the northern part of the basin. The belt of manganese deposits assigned to the Pod"emsk suite by Ustalov [6] can be traced along the western limb of the Porozhinsk syncline and southward along the eastern limb of the small Krivinsk syncline (Fig. 1).

The stratotype of the suite is located to the northeast, in the valley of the Pod"em river, a tributary of the Teya river. Here, it lies with an angular unconformity on metamorphosed deposits of the Sukhopitsk suite and with a disconformity on the Suvorovsk suite. In the lower reaches of the Pod"emsk suite, layers of dolomite, limestone, marl, and siltstone alternate with one another. The middle and upper parts of the suite are composed of sandstones and siltstones with lenses of conglomerate-breccia and gritstone. The total thickness of the suite does not exceed 600 m [5].

In the Vorogovka trough, the stratigraphic position and composition of the Pod"emsk suite is different. It has accumulated a section of unmetamorphosed deposits in unbroken sequence. The underlying deposits here are rhythmically layered grey limestones of the Sukhorechensk suite. Lying on top of these is an horizon of massive oncolitic and stromatolitic dolomites assigned by Ustalov to the lower subsuite of the Pod"emsk suite. In the dolomites near the Vorogovka river, he distinguishes between four members (bottom to top): light-grey massive dolomite (30-100 m); brecciated grey dolomite and conglomerate-breccias (45-190 m); light-grey dolomite containing sandy-silt-size quartz (170-200 m); variegated limestone and dolomite with argillite beds (80-90 m). These members cannot be traced to the north, however. Numerous drill holes here penetrate massive or, less commonly, light-grey, locally rosy, algal dolomite. Thin layers of greenish-grey and cherry-red dolomitic siltstone and marl and lenses of dolomitic limestone are found within the algal dolomite. The thickness of the dolomitic subsuite in the Porozhnaya river basin ranges from 300-400 m and averages 360 m.

The volcanogenic – terrigenous manganiferous first member lies upon dolomite with a sharp contact but without an apparent stratigraphic gap. V. V. Ustalov classifies it as the first member of the upper subsuite of the Pod"emsk suite. The lithological features of the member are discussed below in detail. We will only note here that its composition and thickness are highly variable.

With gradual transitions expressed by increased contents of quartz and chalcedony, the manganiferous member is replaced by a second, tuff-silicite member. It is composed of microlaminated, strongly fissured rock with a color ranging from brownish-white to black; this is a result of different degrees of leaching of the rock by fissure-stratal waters. A chalcedony-quartz aggregate makes up the bulk of the tuff-silicite and tuff-phthanite. There are bands of silt-size quartz

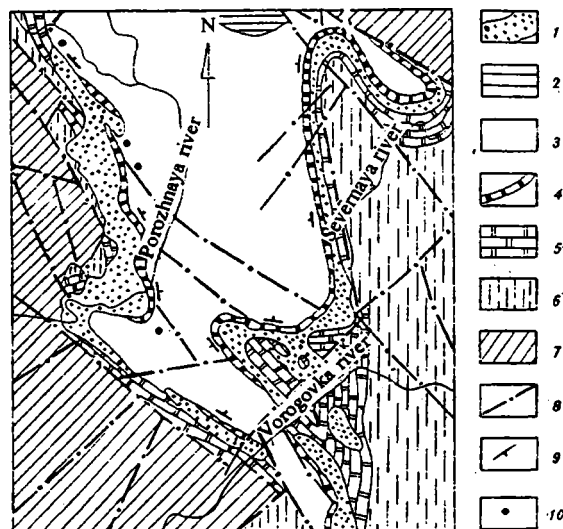


Fig. 1. Geological map of the Porozhinsk manganese district (based on materials from the Krasnoyarskgeologiya Company): 1) deposits of karst depressions (Mesozoic and Cenozoic); 2) deposits of the Lower Cambrian; 3) deposits of the Upper Riphean-Vendian, undifferentiated (upper portions of the Pod'emsk and Nemhansk suites); 4) tuff-phthanite of the Pod'emsk suite; 5) dolomite of the lower reaches of the Pod'emsk suite; 6) deposits of lower part of the orogenic complex of the Upper Riphean (Sukorechensk, Mutninsk, and Severorechensk suites); 7) deposits of a geosynclinal complex (Sukhopitsk series of the Upper Riphean); 8) dislocations; 9) attitude of Pod'emsk suite; 10) drill holes penetrating complete sections of the manganiferous member, slightly affected by weathering.

within it. Some grains have angular, chipped contours suggesting a pyroclastic origin. Fragments of feldspar, acidic effusive rock, and mica flakes are less common. The thickness of the tuff-silicite (second) member of the Upper Pod'emsk suite equals 120-195 m.

Siltstones and tuff-siltstones of the third member lie on tuff-silicates. These are finely layered and foliaceous-layered rocks of gray and greenish-gray color, often calcareous. Flakes of muscovite and biotite are noticeable on bedding planes. There are lenses of calcareous-chalcedony rocks and manganiferous arenaceous limestones. The thickness of the member is 140-170 m.

These grade into a carbonate-tufogene-terrigenous Quaternary member 120-140 m thick. This member is composed of light-gray oncolitic, stromatolithic, and microgranular limestones, calcareous siltstones, and tuff-siltstones. Some of the limestones varieties are manganiferous (MnO to 7%).

The fifth member of the suite has been penetrated near the core of the Porozhinsk syncline, where it is composed of gray and greenish-gray siltstones and fine-grained sandstones and lenses of stromatolithic limestones. The thickness of the member is 215-240 m.

Deposits of the sixth member, represented by arenaceous limestones and fine-grained calcareous sandstones crown the section of the Pod'emsk suite in the Porozhnaya river basin. The total thickness of the Pod'emsk suite in the Porozhinsk syncline is 1100-1350 m [6]. In an unbroken succession, the syncline accumulated the more coarsely fragmental Nemchansk suite of comparable thickness.

Features of composition, degrees of lithification, and modes of occurrence of deposits of the Upper Riphean suggest that the Vorogovka trough belongs to the orogenic stage in the development of the Yenisey fold structure of the Baikal tectonic cycle [5]. From bottom to top, the formation series is comprised of a lower molasse, carbonate-terrigenous carbonate, carbonate-volcanogenic-terrigenous formations, and an upper molasse. The carbonate (limestone-dolomite) formation corresponds to the maximum Late Riphean transgression (inundation), after which a stage of differentiation

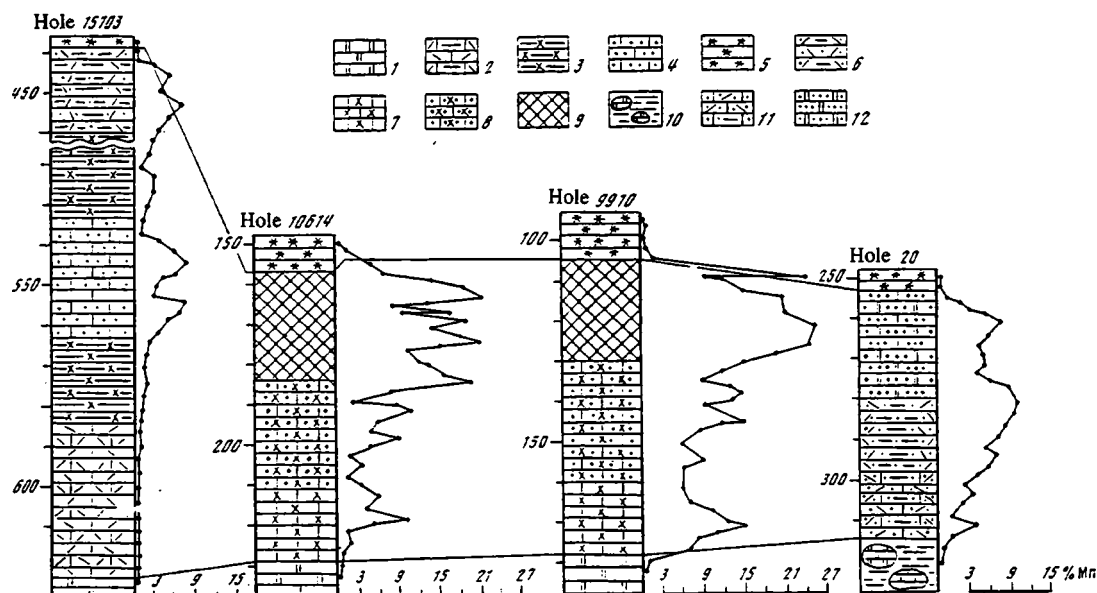


Fig. 2. Lithological columns of manganiferous member of the Pod'emsk suite (north to the left): 1) dolomite; 2) dolomitic limestone with admixtures of pyroclastic material; 3) siliceous aleuropelitic tuffite; 4) rhodochrosite tuff-sandstone; 5) tuff-phthanite; 6) silty crystal-vitroclastic tuff and tuffites 7) siliceous-carbonate aleuropelitic tuffite; 8) same, silty-sandy; 9) carbonate manganiferous ore; 10) fragments of dolomite in powdery mass; 11) silty-sandy carbonate tuffite; 12) rhodochrosite-dolomite rock.

was accompanied by rhyolitic volcanism on nearby land. During Nemchanian time (Vendian), a regression took place, after which the trough closed. We can identify the manganiferous paragenation on the basis of the carbonate–volcanogenic–terrigenous formation associated with the stage of differentiation. In terms of composition, it can be assigned to a group of carbonaceous manganiferous parageneses characteristic of the Late Precambrian [1].

The geological structure of the Porozhinsk manganese-ore district has been variously interpreted. Mstislavskiy and Potkonen [3] assume that the essentially manganite and rhodochrosite lenticular-bedded mineralization is hydrothermal-sedimentary. They believe that it is associated with taphrogenesis complicating the evolution of the Vorogovka trough and resulting in a mosaic-block structure for the latter. At the base of the ore member, they diagnose ore breccias from a volcanotectonic breakup. Accumulations of black manganese and manganosite formed at nodes of fault intersection. Karst developed at juncture zones.

Both the authors and Pasashnikova [4, 7], who has provided geological leadership for the survey work in the region, suggest that ore-localizing structures formed in several stages manganocalcite- and rhodochrosite-bearing tuffites, tuffs, carbonaceous siltstones, volcanomictic sandstones, and lenses of ore of the same composition formed during the sedimentary-diagenetic stage. During this stage, epi-Baikalian tectogenesis created the fold-block structure of the trough and hydrothermal activity appeared at a small scale. Manganiferous deposits underwent catagenesis accompanied by partial redistribution of manganese carbonates. During the Mesozoic–Cenozoic stage, intensive hypergenesis took place. During the Middle Jurassic–Neocomian, deep contact-karst depressions formed. Deposits in these depressions included lenses of oxide ores (manganite, cryptomelane, pyrolusite, todorokite, etc.) and, more rarely, carbonate ores. During the Cenozoic, widespread chemical weathering (a large portion of which was neolluvial) and karst processes in the roof portion of the dolomites underlying the ore member took place. We interpret the ore breccias as karst-subsidence formations.

In order to interpret the conditions of lithogenesis of the ore member (the sedimentary-diagenetic stage), let us consider the portion of the member weakly affected by hypergenesis. The member can be traced for 16 km on the basis of drilling profiles. In the southern part of the ore district, its thickness varies in thickness from 23 m (Hole 29) to 102 m (Hole 51) and changes abruptly over a distance of 1–1.5 km.

At Hole 51, dark-grey banded-layered siliceous siltstones (18 m) lie on light-grey dolomites and are overlain by grey and dark-grey clayey and micaceous tuff-siltstones (30.5 m). These rocks lighten upward and have been converted

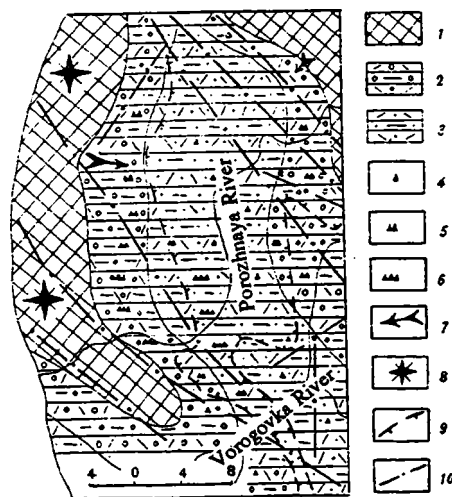


Fig. 3. Lithological-paleogeographic map corresponding to the beginning of the Upper Pod'emian (Late Riphean): 1) eroding land; 2) shelf deposits; 3) deepwater deposits; 4-6) slight, moderate, and substantial deposition of manganese carbonates; 7) direction of drift of sedimentary material and manganese compounds; 8) volcanos; 9) zone of present-day distribution of deposits; 10) faults affecting the depositional environment.

into saprolites (16 m). Monotone foliaceous-layered grey and dark-grey mangiferous tuff-siltstones can be traced next (43 m). According to core sampling data, the content of manganese in these deposits varies from parts of a percent to 4.8% in tuff-siltstones overlying tuff-sandstones.

In the section of Hole 20, located 1 km to the northwest (Fig. 2), the roof portion of the dolomite has been altered by hypergenesis. Grey and dark-grey foliaceous-layered mangiferous silty-sandy tuffite (argillized in the lower reaches of the bed) lies on white carbonate clay with gravelly and rubbly dolomite. This tuffite is overlain by outwardly similar aleurolitic layered tuffite rich in manganese (up to 9.6%). One can then trace a grey, greenish grey rhodochrosite-dolomite rock formed as a result of diagenetic carbonatization of psammitic tuff.

Drill holes located to the northwest, along the bed of the Porozhnaya, penetrated epigenetic oxide ores confined to the middle and lower parts of a hypergenetically altered ore member 34-41.5-m thick (Holes 12, 60, etc.). Mangiferous deposits slightly affected by weathering were penetrated at shallow depths 9 km northwest of Hole 9910. Gray and greenish-gray foliaceous-layered siliceous-rhodochrosite tuff-siltstones grading into dark-gray siliceous-rhodochrosite silty-sandy tuffite lies upon light-gray massive dolomite. Above, the hole penetrates a thick (26 m) lens of manganese carbonate ore represented by an indistinctly layered aggregate exhibiting psammitic structure and giving the appearance of tuff-sandstone. The section penetrated by the nearby hole 10614 is completely analogous.

To the north, the structure of the ore member changes substantially. Thus, the thickness of the member at Hole 15703 is markedly greater due to silty tuffite, but the content of manganese is lower, not exceeding 7-8% in the middle of the member.

In the lower part of the member, aleurolitic foliaceous-layered dark-gray tuffite containing variable quantities of quartz-chalcedony and dolomite cement predominate (see Fig. 2). Pyroclastic rock consisting of quartz, less commonly glass and feldspar, are recognizable among the fragments. These minerals are corroded and replaced by the cementing material. Microglobules of quartz embedded in a brown clayey mass are distinguishable in the groundmass of the latter at high magnifications. Larger dimensions and quantities of pyroclastics (silty and silty-sandy tuffite and tuff) occur locally in the middle of the member. The tuffs also exhibit a high degree of geochemical reworking, with replacement by a rhodochrosite-dolomite association; high-manganese rocks and carbonate ores (>10% Mn) have formed where this predominates. These rocks and ores contain polygonal grains and grain fragments composed of rhodochrosite, mangandolomite

and dolomite as well as spherulites of ellipsoid, pear-shaped, globular, and dumbbell shapes. Their sizes range from 0.07 to 0.6 mm. A portion of these grains are apparently lithoclasts replaced by carbonate minerals, with a predominance of nodules formed from one type of them.

The tuff-sandstone found everywhere in the roof portion of the member forms a bed of variable thickness (from 0.7 m in Hole 28 to 26 m in Hole 9910-10614). The tuff-sandstone has also been subject to intensive geochemical reworking and, in places, is replaced by ore carbonates.

The content of manganese along the section varies abruptly, showing a tendency to increase in middle and upper parts of the member with epigenetic accumulation of the dolomite-rhodochrosite association.

The data presented suggest that conditions for the formation of the manganese member began during the Upper Pod'emian. At the beginning of the stage of tectonic differentiation of the Vorogovka trough, a structural reorganization took place. The large shallow water basin in which dolomites had been deposited was transformed into a smaller basin with islands and rugged bottom topography. Volcanos were active on land and supplied pyroclastic material of rhyolitic composition to the sea. We suggest that volcanic edifices existed above the granite stocks of the Kutukas-Berezovsk uplift (Fig. 3). Pyroclastics settling into the water were subjected to geochemical reworking. The depth of the sea was substantial (the bathyal zone), as indicated by parallel banded layering, elevated content of organics, and fine dissemination of pyrite. At the same time, the sediment accumulated above the carbonate compensation depth.

The variability in the thickness of volcanogenic-terrigenous deposits can be explained by irregularities of the basin floor adjacent to the continental slope. The presence of barriers caused different degrees of infilling of the basin. Manganese entered the sediment in the form of suspensions and accumulated in more substantial quantities where dilution with terrigenous and pyroclastic material was relatively minor. The sources of ore material were located toward the western coast. Muddy waters dissolved ore suspensions and resulted in convective-diffusive redistribution of manganese which had bonded to sediment when the saturation limit for carbonate phases (rhodochrosite, mangandolomite, and mangancalcite) had been reached.

In general, the ore member reflects a single regressive rhythm. As regression continued, the supply of manganese increased. We can assume an initial volcanogenic-sedimentary source of the latter, since deposition of manganiferous sediments and pyroclastics took place simultaneously and both were supplied from the landward side.

Subsequently, another reorganization of paleogeographic conditions took place. The area under consideration rapidly subsided below the level of carbonate compensation, the supply of terrigenous material was almost discontinued, and the sizes of pyroclastics decreased to finely silty-pelitic. Under these conditions, deposition of the tuff-phthanite members took place. These deposits are not manganiferous, since conditions for binding of manganese carbonates to sediment were absent.

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BURIED NODULES IN THE EASTERN EQUATORIAL ZONE OF THE PACIFIC OCEAN

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UDC 552.124.4 (265/266)

The composition and makeup of nodules in Pleistocene deposits have been investigated. Nodule growth rates were calculated on the basis of burial age. The presence of large buried nodules, differences in the structure and composition of the ore matter in surface- and buried nodules, indicate the presence of several nodule-forming stages during the Pleistocene.

Information on buried nodules from the Equatorial Pacific is known from numerous publications [7, 8-10, 16, 19, 24, 30, 36-38, etc.]. Glasby [24], who analyzed nodule distribution patterns in cores from deep water drillholes, noted the dominance of buried concretions in the north Equatorial Pacific. Earlier, Menard [30] mentioned the abundance of nodules in the eastern Equatorial belt.

Drill data indicated that nodules in the Clarion-Clipperton ore province occur in sediments that range in age between Middle Eocene and Pleistocene [16, 24]. Cruises by RZV *Vahine*, *Vityaz'*, *Dmitrii Mendeleev*, *Valdivia*, *Zonne*, and others, led to discoveries of nodules in sediments of Lower Miocene-to-Pleistocene age [9, 16, 36, 37, etc.]. Sediments that contain the nodules are clayey-silica muds, pelagic clays; rarely carbonate muds. Nodules in the far-eastern part of the Equatorial zone (Guatemala and Peru Trenches) occur in Pleistocene-age silica-clay muds [21, 39, 40].

Data regarding nodules and obtained by the 127-mm direct-flow corer UT-73 during the 41st Cruise of R/V *Dmitrii Mendeleev* in the eastern part of the Clarion-Clipperton ore Province and in the Guatemala Trench, are presented (Table 1, Fig. 1).

RESEARCH METHODS

Deposits that enclose nodules were studied by microscope in permanent prepares. Sediment ages were determined by microfossil (diatom and coccolith) analysis and paleomagnetic methods.

Age zonations of tropical bottom deposits are based on a large body of data, employing diatom and nannoplankton information that is based on the irreversible evolution and ecological changes of these taxonomic groups. This is expressed by the presence of certain taxa (UP), extinction of others (UPN), as well as brief flourishing of given forms (UO). The absolute dating of time horizons and their succession were based on paleomagnetic and radiometric geochronology.

Ten diatom-dated time horizons were utilized in our study [1]: 1.56 Ma – UPN *Rhizosolenia praeberaonii* v. *robusta*; 1.0-0.92 Ma – brief occurrence zone of *Rhizosolenia matujamai*; 0.98-0.91 Ma – UO *Mesocena elliptica* [5]; 0.79 – UPN *Mesocena elliptica*; 0.745 Ma – UO *Thalassiosira oestrupii* (small forms); 0.73 Ma – UPN *Nitschia fossilis*; 0.63 Ma – UPN *Nitschia reinholdii*; 0.61-0.62 Ma – UO *Roperia tessalata ovata*; 0.6-0.5 – UPN *Nitschia prolongata* and *Thalassiosira leptopus elliptica*; 0.35-0.3 Ma – UPN *Thalassiosira plicata*; 0.1-0.08 Ma – UPN *Coscinodiscus pseudoincertus*.

Corresponding ages obtained through coccolith dating are the following: 0.07, 0.27, 0.44, and 0.92 Ma [22] was employed in cores from Stations 3875 and 3883 [4]. Paleomagnetic analyses were performed by Yu. Yu. Ivanov; horizon ages were based on work by Mankinen and Dalrymple [29].

The maximum and minimum diameters of each nodule in the cores were measured and the composition of nuclei analyzed. The textural-structural makeup has been investigated and the surface characteristics studied in polished

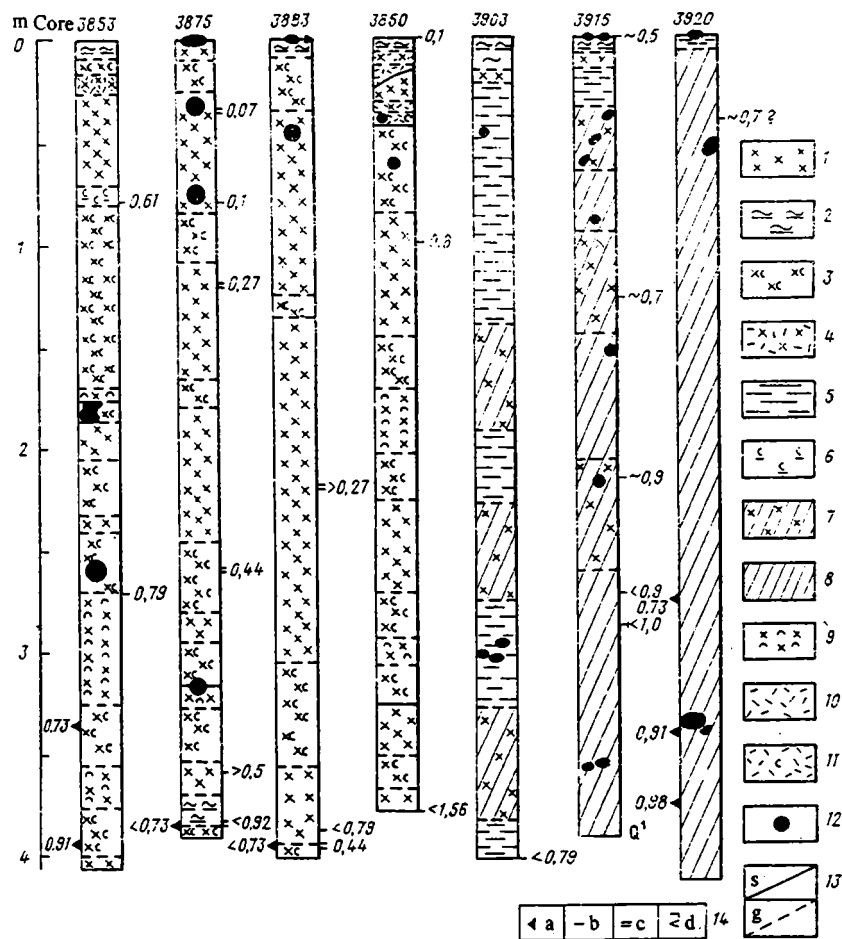


Fig. 1. Lithostratigraphic subdivision of sediment columns, eastern Equatorial Pacific: 1) clayey-radiolarian mud; 2) same, diatom-enriched; 3) same, slightly limey; 4) same, tuffitic; 5) radiolarian-clayey mud; 6) same, slightly limey; 7) eupelagic clay; radiolarian-enriched; 8) eupelagic clay; 9) limey-clayey mud, radiolarian-enriched; 10) tephra; 11) same, slightly limey; 12) Fe-Mn concretions; 13) division boundaries (s - sharp; g - gradual); 14) age-dated; a) paleomagnetic date; b) diatom date; c) coccolith date; d) radiolarian date).

sections. Usually, the nodules were cut into two; one half was used for section preparation and subsequent layer-by-layer mineralogical and chemical analyses. Samples were taken for chemical analyses from the second half. An averaged concretion ore envelope sample was usually analyzed. Individual layers with contrasting reflection characteristics were sampled in the concretions (Table 2 and 3). T. G. Kuz'mina performed total silica analyses by the Carl Zeiss X-ray spectral fluorescence analyzer VRA-20. Fe, Mn, Ni, Co, Cu, Zn, Pb, Mo, V, Li, and Cd analyses were performed by N. N. Zavadskaya, with a Perkin-Elmer atomic absorption spectrophotometer for flame operations.

Nodule growth rates were calculated by the age of the enclosing sediments and Mn/Fe ratios [26, 28]. Employing the generally accepted concept that nodules generally grew on surfaces of sedimentary mud layers or on semiliquid sediment layers, the diagenetic concretions started to develop at the time when the largest of the preexisting-surface nodules was buried. The time of burial and the corresponding time interval of growth of the overlying nodules were calculated with the help of mean sediment deposition rates, recorded for given geological time intervals (Tables 1-4; Fig. 1). Thus, the lower boundary of the Brunhes Event (0.73 Ma) in Core 3920 is located at 275 cm, corresponding to a sedimentation rate of 3.8 mm/ka (Table 1). Nodules occur on the seafloor and in the deposits in the 50-54 cm core interval. By using the mean sedimentation rate that prevailed at the time when the concretions of the 50-54-cm interval were buried; their age is calculated as 0.132 Ma (the overlying 50-cm sediment was deposited at the rate of 3.8 mm/1 ka). However, nodule growth may have been initiated well after the preexisting nodules became buried. This means that the

TABLE 1. Buried Nodules in East Equatorial Pacific

Station No.	Coordinates	Depth	Morphology and concretion size, cm	Morphology and concretion size, cm	Enclosing sediments				
					lithological characteristics	age		deposition rate	
						horizon depth, cm	Ma	thickness of layer, cm	mm/ka
Clarion-Clipperton ore province									
3920	14°39,3' latitude, N 126°00,8' longitude, W	4495	0	fBbr 5×4,5×3,5	Radiolarian-clay muds	30	0,7?	30	~0,43*
					light cinnamon-brown	275	0,73	275	3,8***
					Eupelagic clay, cinnamon-colored, marbled	337	0,91	62	3,4***
					Same, pale yellow	375	0,98	38	5,4***
3916	13°30,0' latitude, N 126°00,6' longitude, W	4560	330-336	fEDbr/s 8×5×4,5	Same				
			340-342	Tbr 2×1,5×0,5	Same				
			0	Ts 4×3×1; 3×2×1	Clayey-radiolarian mud	0	0,5	-	-
					with abundant diatoms	130	0,7	130	6,5*
					cinnamon-colored	215	0,79	85	9,2*
					Eupelagic clay,	270	0,9	55	5,0*
					enriched in radio-	290	>1,0****		
					larians, light-cinnamon-				
					colored, marbled				
					Same				
					Same				
					Eupelagic clay, light-				
		cinnamon colored							
		147-152	lEr 5×3,5×3	Same, cinnamon-colored					
		215-218	lDr 3×4×1,5	Same, enriched in radiolarians					
		353	lFs <2	Same					

3903	13°59,9' latitude, N 112°31,2' longitude, W	3970	45 291-294 300-303	IDr 2 Dr 3 IDr interlayer 2-3 Guatemala trench	Radiolarian-clay mud, cinnamon-colored Same	380	0,79	380	4,8*
3850	6°37,5' latitude, N 93°21,0' longitude, W	3600	30-34 55-62	ITr 3×3,5×0,8 SEr/s 7×6×3	Slightly limey tephra Clayey-radiolarian mud, slightly limey, cinnamon- colored, marbled	0 100	0,1 0,9	— 100	— 1,2*
3853	6°32,2' latitude, N 92°48,5' longitude, W	3595	175-185	ISbr 10×10×11	Clayey-radiolarian mud, slightly limey, light- cinnamon-colored, marbled	80 270	0,6 0,79	— 190	— 10,0
3875	6°00,3' latitude, N 92°48,5' longitude, W	3610	255-266 0 27-34 66-74 310-318	ISbr/s 11×10×10 IEr/s 9×6×4 ISbr 7×7× 7 ISbr 9×8×8 ISbrs 9×9×9	Same, cinnamon-colored Clayey-radiolarian mud, dark-cinnamon-colored Same " Clayey-radiolarian mud, slightly limey, light- cinnamon-colored	35 80 120 270 360	0,07 0,1 0,27 0,44 >0,5	35 45 40 150 90	5*** 15* 2,3** 8,9** 15
3883	6°30,1' latitude, N 93°00,1' longitude, W	3610	0 50-54	EsR/s 4×4×3 Er/s 6×5×4,5 ISr 6×6×6	Clayey-radiolarian Diatom-enriched, dark cinnamon-colored Same	220	0,27	220	8,3**

*Based on diatom analysis data (Fig. 1).

**Based on coccolith spectra.

***Based on paleomagnetic data.

****Based on radiolarian spectra.

TABLE 2. Chemical Composition of Buried Nodules, Clarion-Clipperton Ore Province, Eastern Radiolarian Belt, Pacific Ocean

Station No. and horizon depth, in cm	Nodule morphology	Characteristics of analyzed samples	Al	Si	Mg	K	Ca	Ti	Fe	Mn	Co	Ni	Cu	Zn	Pb	Mo	V	Li	Cd	Mn/Fe
3920-0	Eb s/r	Sample mean	2,59	7,65	1,83	0,81	1,70	0,34	5,75	21,15	0,23	2,32	1,22	0,157	—	—	—	—	—	4,7
50-54	Dr	Same	2,75	8,46	1,83	0,74	1,85	0,35	4,68	27,32	0,27	1,60	1,41	0,154	—	—	—	—	—	5,84
330-336	Dbs/r	"	2,33	7,31	1,80	0,78	1,83	0,65	10,31	27,50	0,35	1,07	0,68	0,10	—	—	—	—	—	2,67
345	Tb s/r	"	4,11	—	1,13	1,08	0,736	0,42	6,55	19,63	0,51	0,84	0,79	0,057	0,096	0,025	0,0159	48	8	3,0
3916-0	Ts	"	—	—	1,87	—	—	0,57	12,84	28,16	0,32	1,11	0,65	0,093	0,073	0,062	0,052	78	8	2,11
0	ITs	"	2,44	11,0	2,03	0,8	1,17	0,66	13,24	19,79	0,25	0,71	0,57	0,060	0,094	0,033	0,047	87	11	1,43
37	Ifs	"	—	—	—	—	—	0,14	18,8	6,73	0,18	0,23	0,25	0,039	0,02	0,014	0,103	49	1	0,37
50-53	Er	"	3,19	11,02	1,88	1,18	1,17	0,31	7,61	25,71	0,24	1,10	0,76	0,081	0,038	0,049	0,49	118	13	3,37
60-62	IDr	"	2,54	7,33	2,08	0,95	1,34	0,38	6,82	31,83	0,26	1,17	0,84	0,076	0,043	0,054	0,053	129	14	4,68
90-92	Ir	"	2,21	6,63	1,97	0,99	1,37	0,3	6,42	34,89	0,23	1,07	1,02	0,093	0,036	0,059	0,065	102	18	5,43
147-153	a IEr	Upper	—	—	1,66	—	—	0,39	6,52	32,03	0,19	0,89	1,02	0,095	0,034	0,049	0,032	347	13	4,9
147-153	b "	Lower	—	—	1,77	—	—	—	4,84	35,30	0,15	1,33	1,06	0,103	0,036	0,047	0,039	328	13	7,3
147-153	c "	Middle	2,24	6,7	1,87	0,89	1,39	0,36	7,61	32,64	0,23	1,23	0,98	0,094	0,040	0,064	0,086	90	16	4,29
215-218	IDr	Same	2,59	8,41	2,06	0,96	1,28	0,37	7,71	28,97	0,30	1,09	0,95	0,070	0,047	0,05	0,069	91	12	3,75
353	"	"	3,93	22,35	1,72	1,83	0,50	0,28	10,08	5,10	0,084	0,13	0,20	0,040	0,018	0,008	0,06	74	1	0,51
3903-45	Dr	"	—	—	1,96	—	—	—	8,50	30,81	0,10	1,25	0,89	0,132	0,028	0,063	0,049	134	21	3,62
300-305	IDr	"	2,23	7,17	1,67	0,66	1,20	0,37	8,92	26,62	0,13	0,79	0,77	0,075	—	—	—	—	—	2,99

Note. Li and Cd content in $10^{-4}\%$; other elements, in percentages; station numbers: figures in heavy print.

TABLE 3. Chemical Composition of Buried Nodules in Guatemala Basin, Eastern Radiolarian Belt, Pacific Ocean

Station No. and horizon depth, in cm	Nodule morphology	Definition of analyzed sample	Al	Mg	K	Na	Ca	Ti	Fe	Mn	Co	Ni	Cu	Zn	Pb	Mo	V	Li	Cd	Mn/Fe
3850																				
30-34	Tr	Sample mean	2,08	2,19	0,60	1,57	1,35	0,14	5,94	33,69	0,086	1,06	0,532	0,228	0,006	--	0,052	15	9	5,7
52-60	Se	"	--	--	--	--	--	--	3,28	43,64	0,03	0,795	0,46	0,228	--	--	--	--	--	13,3
3853																				
175-185	ISb	"	1,35	--	--	--	1,29	--	1,29	50,77	0,05	0,704	0,21	0,153	--	--	--	--	--	39,75
255-266	Isb	"	1,53	--	--	--	1,17	--	1,66	47,76	0,062	0,88	0,226	0,174	--	--	--	--	--	28,7
255-266	"	Upper	1,85	--	--	--	1,45	--	2,53	40,36	0,078	1,36	0,351	0,199	--	--	--	--	--	15,95
255-266	"	Lower	1,15	--	--	--	1,23	--	0,87	50,45	0,05	0,51	0,353	0,183	--	--	--	--	--	57,9
3875-0	IE	Middle sample	1,46	1,93	0,66	3,03	1,42	0,07	1,68	46,82	0,018	0,61	0,25	0,21	0,004	0,051	0,036	--	--	27,86
		Upper	1,58	--	--	--	1,52	--	1,84	45,16	0,012	0,759	0,30	0,275	--	--	--	--	--	24,5
		Lower	0,65	--	--	--	1,35	--	0,58	52,22	0,006	0,148	0,13	0,067	--	--	--	--	--	90,0
		MD	0,90	0,94	0,53	3,00	0,86	0,03	0,94	44,98	0,005	0,339	0,144	0,126	0,007	0,052	0,0016	13	39	48,0
		MPD	1,98	1,16	0,63	2,56	0,81	0,09	2,11	38,46	0,006	0,713	0,338	0,225	0,011	0,048	0,0013	11	29	18,2
		MD	1,06	1,4	0,55	3,35	0,81	0,027	0,97	41,05	0,009	0,62	0,35	0,29	0,019	0,050	0,0128	14	33	42,2
27-34	Sbr	Middle sample	--	2,20	--	--	--	0,12	2,55	43,05	0,03	0,72	0,38	0,155	0,007	0,021	0,046	21	2	16,9
24-27	"	MD+M	1,22	1,24	0,66	1,95	0,767	0,05	1,17	43,9	0,009	0,546	0,304	0,105	0,007	0,018	0,021	16	9	37,9
27-34	"	MPD	1,86	1,22	0,63	1,88	0,67	0,10	2,11	37,91	0,009	0,71	0,353	0,155	0,007	0,018	0,0165	2	9	18,0
66-74	Isb	Middle sample	1,78	--	--	--	1,18	--	2,11	44,99	0,031	0,75	0,162	0,142	--	--	--	--	--	21,32
66-74	"	MD+M	1,99	2,72	0,77	2,00	1,27	0,11	2,65	41,12	0,029	0,689	0,168	0,167	0,007	0,018	0,0148	18	14	19,3
66-74	"	MPD	1,53	2,66	0,66	1,77	1,18	0,07	2,16	34,56	0,03	0,669	0,168	0,164	0,003	0,017	0,046	16	9	11,3
310-318	Isb	Middle sample	1,37	--	--	--	1,04	--	1,78	50,0	0,053	0,75	0,127	0,16	--	--	--	--	--	28,09
310-318	"	MD+M	1,20	3,02	0,74	2,04	1,16	0,04	1,46	46,42	0,064	0,669	0,187	0,188	0,0013	0,02	0,061	12	11	47,9
310-318	"	MPD	1,53	3,02	0,80	2,16	1,26	0,09	2,22	41,37	0,069	0,644	0,157	0,168	0,009	0,016	0,063	11	13	18,6
3883-0	Es	Middle sample	--	2,29	--	--	--	0,07	1,58	49,17	0,02	0,47	0,27	0,186	0,004	--	--	8	5	31,0
50-54	Is	Same	1,38	2,25	0,62	1,85	1,11	0,065	2,44	46,03	0,025	0,68	0,343	0,171	0,004	--	0,048	15	7	18,9
50-54	"	MD+M	1,11	2,22	0,62	1,93	1,05	0,03	1,24	50,02	0,023	0,62	0,364	0,0154	0,004	--	0,052	14	6	40,3
50-54	"	MPD	1,65	2,28	0,62	1,77	1,18	0,10	3,64	42,04	0,27	0,77	0,422	0,188	0,004	--	0,044	16	9	11,5

Note. Li and Cd content in 10^{-4} %: the rest, in percentages.

TABLE 4. Nodule Growth Rates

Station No.	Horizon, cm	Age of nodules burial in the enclosing deposits, Ma	Duration of nodule growth, Ma	Nodule growth rate, mm/Ma	
				in terms of sediment age	in terms of Mn/Fe ratio
3920	0	—	0,132	189 } 27	13
	50—54	0,132	0,757		20
	330—336	0,889	—		—
3916	0	—	0,577	5	2,8
	50—90	0,577	0,145	42	7—13,6
	147—152	0,722	0,068	360 } 87	9
	215—218	0,79	>0,210		7,8
	353—355	>1,0	—		—
3903	45—47	0,094	0,531	18,4	6,8
	291—305	0,625	—	—	—
3850	30—34	0,340	0,200	87	14,6
	55—62	0,540	—	—	107
3853	175—185	0,695	0,08	690	487
	255—266	0,775	—	—	279
3875	0	—	0,054	830	252
	27—34	0,054	0,037	946	106
	66—74	0,091	0,376	119	176
	310—318	0,467	—	—	251
3835	0	—	0,062	480	304
	50—54	0,062	—	—	124

burial age provides a maximum value for the duration of nodule growth and, correspondingly, a minimum value for calculating the time interval of ore envelope growth on the nodules.

A systematic relationship exists between Mn/Fe ratios and nodule growth rates [26]. By employing the $^{230}\text{Th}/^{10}\text{Be}$ isotope ratios, Lyle [28] recommended the following equation in calculating nodule growth rates:

$$S = 16,0 \text{ Mn/Fe}^2 + 0,448,$$

where S equals the nodule growth rate, in mm/Ma. Growth rates in Table 4 were based on this equation.

DISCUSSION OF DATA

Data collected by the 41st Cruise of R/V *Dmitrii Mendeleev* were obtained from the Clarion—Clipperton ore province. They showed that buried nodules occurred in Pleistocene deposits at three stations (3920, 3916 and 3903 (Fig. 1, Table 1). Intermediate-size and small (2-6 cm) nodules predominated among the buried ones. Often they were partially decomposed. In terms of surface, textural-structural, and compositional characteristics, these are mostly diagenetic and depositional-diagenetic nodules, with typical 2.5-5.8 Mn/Fe ratios, increased Ni, Cu, and Zn content and high (0.8-1.0) Cu/Ni values.

Nodules at Station 3920 (Table 1; Figs. 1 and 2) occur on the surface of radiolarian-clay muds. They occurred in eupelagic clay beds at three levels, to a depth of 342 cm. Paleomagnetic data indicate that the sediments that contain nodules postdate 0.91 Ma (Jaramillo Event). Nodules on nodule-group surfaces, as well as in the 50-54-cm and 330-336-cm depth intervals have disintegrated. Surface nodules, maximum 5 cm in diameter, are ellipsoidal-botryoidal, with an asymmetrical surface texture. The nodule side and lower surface are globular; the highest projection, smooth. Asymmetry is also displayed in thickness differences of the ore envelope (the lower portion is thicker). Similar asymmetry was displayed by buried nodules (horizons 50-54 and 330-336 cm). Its original diameter, judging from the thickness of the ore envelope in the nondecomposed portion of the nodule was 6.5-7 cm. Clay, largely replaced by ore matter served as nuclei of the surface nodules (Fig. 2a).

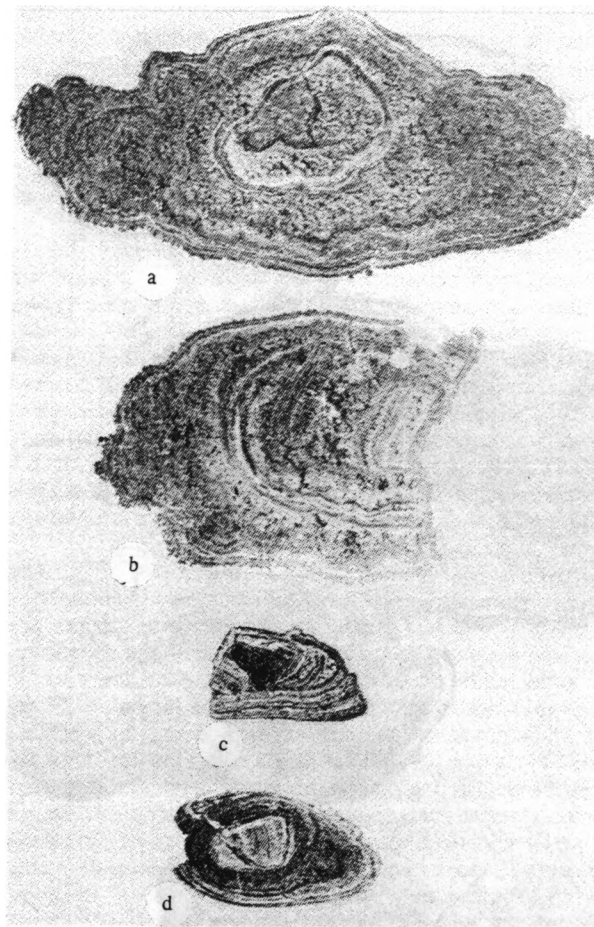


Fig. 2. Interior nodule structures, Stations 3920 and 3903. Polished sections: a) Station 3920, surface horizon, $1.8\times$ magnification; b) same, 50-54-cm horizon; $1.2\times$ magnification; c) Station 3903, surface horizon, $1.1\times$ magnification; d) same, 303-305-cm horizon.

Irregular-discoidal, partly decomposed nodules with globular surface texture occurred at 50-54 cm. Nodule diameters reached 5 cm. Fragments of concretions, formed in earlier generations formed the nuclei (Fig. 2a). Based on the thickness of the outer ore envelope that preserved parts of the nodule, the nodule's original diameter must have been 6 cm.

Discoidal-ellipsoidal nodules from 330-336 cm depth, $8 \times 5 \times 4.5$ cm in size were partially decomposed and possessed asymmetrical surfaces (smooth upper surface), asymmetric composition and ore envelope thickness patterns. Based on the thickness of the preserved portion and the ore crust, its maximum diameter had been 10-11 cm. A small, irregularly shaped tabular nodule at 340-342 cm may have been the buried and decomposed fragment of an ore envelope from a nodule that is located at a depth of 330-336 cm.

In terms of the Mn/Fe ratio and concentration of lesser elements, two nodule groups occurred at Station 3920. High (4.7-5.84%) Mn/Fe ratios dominate in the upper interval (between 0 and 50-54 cm). The Ni and Cu total ranges between 2.54-3.0%. At lower levels, between 330-336 cm and 340-342 cm, the Mn/Fe ratio drops to 2.6-3.0 and the combined Ni + Cu content declines to 1.60-1.75% (Table 2).

In Core 3920, minimal growth rates were calculated from the burial age. The nodules on the surface and in the 50-54-cm horizon may have started to develop well after burial of the preexisting sediment layer. The time elapsed since burial of horizon 50-54 cm was 0.132 Ma; formation of horizon 50-54 cm took 0.889 Ma. Based on the nodule size (calculation based on maximum concretion radius; Table 1), and the time interval of their growth, the nodule growth rate was calculated as 189 mm/Ma; that of horizon 50-54 cm, 33 mm/Ma 0.889 Ma. Mn/Fe ratios show higher rates for

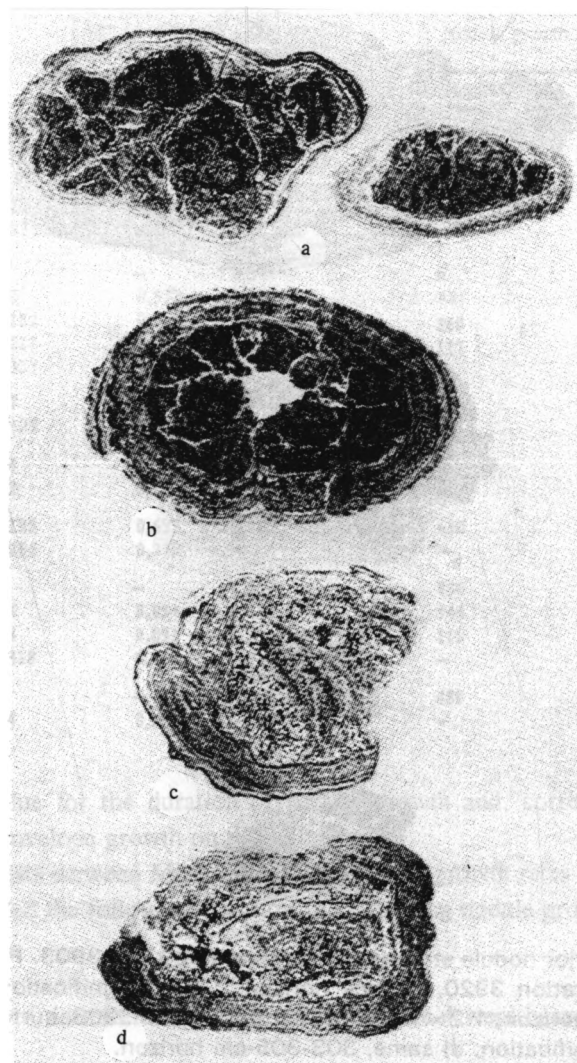


Fig. 3. Interior nodule structures, Station 3916 (Photography by T. Yu. Uspenskaya). Polished section, $1.4\times$ magnification; a) surface horizon; b) 50-53-cm horizon; c) 147-152-cm horizon; d) 215-220-cm horizon.

surface nodules but also good agreement with nodule data in horizon 50-54 cm (13 mm/Ma, respectively, 20 mm/Ma; Table 4). However, judging from similarities in chemical composition between the two nodules, a slight increase in the original size seems to have taken place prior to decomposition of the surface nodule (6.5-7 cm vs 6 cm, in horizon 50-54 cm). Their formation probably was simultaneous after burial of the large nodules in horizon 330-336 cm. A 27 mm/Ma growth rate would thus be assumed for the two nodules. This figure agrees well with values derived from the Mn/Fe ratio. Subsequent burial of concretions in horizon 50-54 cm may have been associated with bioturbation and bottom current activity. Similar processes apparently also affected the core of Station 3916.

Nodules occur both on the surface and in a number of sediment horizons at station 3916. Surface nodules overlie clayey-radiolarian muds that include a diatom flora, dated c. 0.5 Ma (Table 1). Downward, the nodules are associated with eupelagic clays that postdate 0.79 Ma (Fig. 1). A layer, composed of angular clay platelets occurs between 353-355 cm with a thin ore crust.

Radiolarian studies by S. B. Kruglikovaya indicated that the enclosing eupelagic clays below 290 cm predate 1 Ma. The structural and surface features, the nature of nodule nucleus and the composition within the available core length marked the existence of three development stages. Surface nodules are lump-shaped, finely laminated (TKS-type), with a

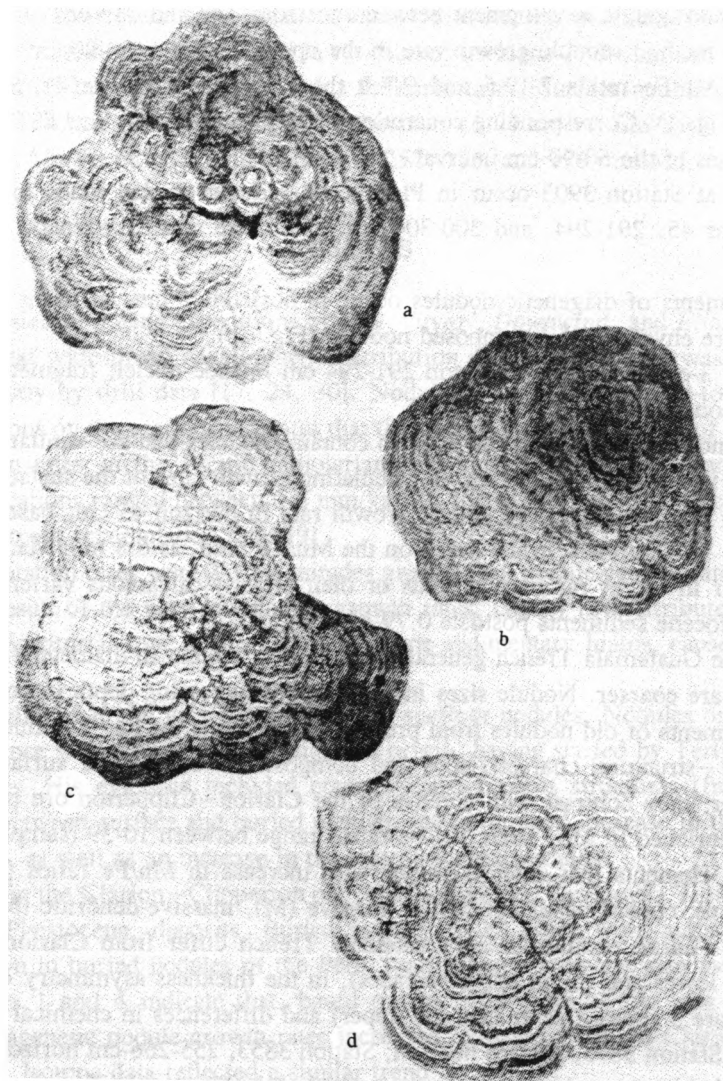


Fig. 4. Interior nodule structures, Station 3875. Polished sections, $1.4\times$ magnification; a) surface horizon; b) 27-34-cm horizon; c) 66-74-cm horizon; d) 310-318-cm horizon.

thickness-asymmetry of the 1-4-mm-thick ore envelope (Fig. 3). These typically sedimentary concretions have low (1.43-2.11) Mn/Fe ratios and 1.28-1.76; low combined Ni and Cu content. The Ti and Pb content is relatively high (Table 2). A compressed fragment of dense clay, with thin iron-enriched crust occurred at 37-cm depth.

Thick-laminated diagenetic concretions with globular surfaces occur in downward direction, within the 50-218-cm interval (horizons 50-53, 60-62, 90-92, 147-153, 215-218 cm). These are enriched in Mn and trace elements (Ni, Cu, Zn), with Mn/Fe ratios of 3.37-5.43 (Table 2). However, nodules between 50-90 cm differ from nuclei of lower (147-218 cm) concretion horizons (Table 1, Fig. 3). Nodules of horizon 50-53 cm are ellipsoidal, of $4 \times 3 \times 2$ cm dimension. Large clayey nuclei and 5-6-mm-thick ore envelope occur. Small (2-2.5 cm) decomposed concretions with clay nuclei relicts occur in horizons 60-62 and 90-92 cm. Ellipsoidal and discoidal nodules in horizons 147-152 cm and 215-216 cm, respectively, possess nuclei that are fragments of diagenetic nodules of earlier generations.

Sedimentary-type concretion embryos occur at 353-cm depth. These are clay platelets with millimeter-thick ore film (Mn/Fe = 0.51). Based on the age of the enclosing sediments, nodules of horizon 50 cm were buried 0.577 Ma ago and C-type surface nodules (Table 4) have been growing since. The ore envelope growth rate was 5.77 mm/Ma; 2.8 mm/Ma, when calculated by the Mn/Fe ratio (Table 4). In calculating ages and growth rates, data from nodules, grouped according to nucleus characteristics, were combined between 50 and 90 cm, and 147-218 cm (horizon 147-152 cm and

215-218 cm). A narrow time gap was assumed between the start of nodule growth in the upper horizon after burial of nodules in horizon 147 cm, and nodule development between horizons 147 and 215 cm (an interval with ore film coated clay platelets). Based on this method, nodule growth rate in the upper horizons (50-90 cm) was 42 mm/Ma, in the lower (147-215 cm), 87 mm/Ma. Mn/Fe ratios 7-13.6 and 9-7.8 the respectively (Table 4). Sediments in horizon 353 cm significantly predate 1 Ma (Fig. 1). Corresponding concretion-growth rates at 147- and 215-cm depths were about half as great; approximately the values of the 50-90-cm interval.

Nodules in the core at Station 3903 occur in Pleistocene radiolarian-clay muds that are younger than 0.79 Ma (Table 1, Fig. 1) in horizons 45, 291-294, and 300-304 cm. These are small diagenetic nodules of maximum 2-3-cm diameter.

Less than 2-cm fragments of diagenetic nodules occur in horizon 45 cm. In terms of structure and composition, nodule nuclei resemble the ore envelope of decomposed nodules (Fig. 2b).

Discoidal nodules of 3-cm diameter in horizon 291-294 cm include nuclei; fragments of a diagenetic nodule that formed in an earlier generation.

A layer of discoidal nodules in horizon 300-305 cm contains nodules that are similar in composition, structure and character of nucleus to those in horizon 291-294 cm. In calculating growth rates of the surface nodules, the time factor was based on combination of the two horizons. The nodule growth rate in horizon 45 cm, based on the age of the enclosing sediments, was calculated as 18.4 mm/Ma; when based on the Mn/Fe ratio, as 6.8 mm/Ma.

Buried nodules occur in clayey-radiolarian muds or their slightly calcareous varieties in the southern Guatemala Trench. The enclosing Pleistocene sediments postdate 0.79 Ma (Fig. 1).

Buried nodules in the Guatemala Trench generally are larger, >6 cm in diameter (maximum 11 cm). At Stations 3853 and 3875, all nodules are coarser. Nodule sizes increased with increasing depth in cores 3850 and 3883 (Table 1). The nodule nuclei were fragments of old nodules from previous generations, or Fe-Mn microconcretions.

In terms of textural-structural characteristics and composition, buried and surface nodules are similar in the Guatemala Trench. However, they contrast with nodules of the Clarion-Clipperton ore province (Fig. 2-4). Those are diagenetic nodules, greatly enriched in Mn. Their Mn/Fe ratios range between 10-39 (sample average, Table 3), with low concentrations of the lesser elements that decline further with increase in Mn/Fe ratios [12, 20]. The nodules display thick-laminated cyclic patterns, consisting of alternating massive (M), massive-dendritic (MD) and massive-mottled-dendritic laminae (Fig. 4) [15]. Large nodules in the Guatemala Trench differ from Clarion-Clipperton nodules in their surface asymmetry (smooth lower and rough upper surfaces), in the thickness asymmetry of the ore envelope (the lower part of the nodules usually are 3 times greater than the upper) and differences in chemical composition (Table 4, Station 3916, 147-153-cm horizon; Station 3875, surface horizon; Station 3853, 255-266-cm horizon).

These nodules included asymmetrical concretions, typical of the Guatemala Trench and nodules with either entirely smooth or entirely rough surfaces. Nodules with variable surface textures in one of the cores were influenced by the disturbance of detrital-feeding in faunal organisms on the sediment surface and during growth.

Guatemala Basin nodules differ significantly in their growth rates. Based both on enclosing deposit ages and Mn/Fe ratios (Table 4), nodule growth rates increased. Horizon 30-34 cm at Station 3850 is an exception. A small, cake-shaped nodule possessed a globular surface texture and interior structure, typical of diagenetic nodules of the Equatorial belt. Massive-dendritic (MD) layers alternated with TSD-layers. The Mn-Fe content was also characteristic. However, the combined Ni + Cu value (1.5%) was lower (Table 4). A large (7 × 6 × 5 cm) ellipsoidal-spherical nodule with composition and structure, typical of Guatemala Trench nodules occurred lower in the core, in horizon 52-60 cm. Nodule growth rate based on the age of the sediments that enclose horizon 30-34 cm, was calculated as 79 mm/Ma; based on the Mn/Fe ratio, as 14.6 mm/Ma (Table 4).

Growth rates turned out to be significantly higher in horizon 55-60 cm when calculated by the Fe/Mn ratio method (107 mm/Ma). Similarly, values increased (120-487 mm/Ma) in all surface and buried nodules in the Guatemala Trench under study.

Buried spherical-botryoidal nodules, of maximum 11-cm diameter, in core of Station 3853 occurred in clayey-radiolarian oozes in horizons 175-185 cm 255-266 cm. The oozes postdated 0.79 Ma. Based on sedimentation rates (Table 1), the burial time of the lower (255-266 cm) horizon and the probable start of nodule growth in the overlying horizon was calculated as 0.775 Ma; that of the upper horizon (175-185 cm), as 0.695 Ma. Nodule growth in the upper horizon lasted for maximum 80 ka. The minimum rate was 690 mm/Ma; approximately the value (487 mm/Ma), obtained for the same nodule by the Mn/Fe ratio.

Large botryoidal-spherical and spherical-ellipsoidal botryoidal nodules, of 7-9-cm maximum diameter occur at Station 3875 on surfaces of three Pleistocene radiolarian-clay mud horizons that postdate 0.5 Ma. Minimum growth rates were calculated for nodules on the sediment surface and in two buried horizons 27-34 cm and 66-74 cm (Table 4). Nodule growth may have started well after burial of the Dreexistina sediment layer. Surface nodule growth rates were calculated as 830 mm/Ma for horizon 27-34 cm and 946 mm/Ma for horizon 66-74 cm on the basis of the age of the enclosing deposits and the time of burial. Mn/Fe ratios produced rates of 252, 106, and 176 mm/Ma, respectively.

RESULTS

Results from cores, collected during expeditions by R/V *Vityaz'*, *Downwind*, and *Elitanin* etc. in the Pacific has a long time ago convinced us that widespread development distribution of the concretions was not restricted Recent times [10]. This was confirmed mostly by drill data [17, 24, 30]. Nodules occurred in Eocene-to-Pleistocene sediments. The distribution pattern of concretions on the seafloor indicates that they formed mostly at times of low sedimentation rates [10, 18, 33]. The mean deposition rates in the Pacific radiolarian belt ranged between 1-3 mm/ka [6]. Specific sediment accumulation rates at given locations ranged between 0-6 mm/ka [8]. Increased nodule production was associated with 1-2 mm/ka, often < 1 mm/ka sediment accumulation rates [9].

Even the slow sedimentation rates were 2-3 magnitudes greater than nodule growth rates (a few millimeters — Ma; Tables 1 and 3) [2]. As the result of the very low nodule growth rates, Glasby [24] attributed the start of their development in the northern Pacific Equatorial zone to the Lower Miocene sedimentary hiatus, caused by intensive, fast-flowing Antarctic bottom currents.

Stackelberg [36-38] believed in the very old age of most seafloor nodules. Nodules on the sediment surface in the Clarion—Clipperton ore province were considered "foreign" in origin, having started by Tertiary (oligocene through Late Pliocene) sedimentary hiatuses. His evidence included compositional nucleus structures (fragments of the depositional nodule category) similarities between surface and buried nodules and an upward increase in the size of ore envelopes that surround nodules in drillcores, as well as an increase in the size of surface nodules.

In cores, investigated in the Clarion—Clipperton ore province and the Guatemala Trench, buried nodules occurred in Pleistocene, mostly Late Pleistocene, deposits. Buried nodules in all cores were associated with erosional unconformities. This was first shown in buried nodules of the Peru Trench where they are also associated with silica-enriched Pleistocene oozes [40]. Tables 1 and 4 indicate that, based on the calculation of the ages of burial and Mn/Fe ratio, sediment accumulation and diagenetic nodule growth rates increased eastward from the Clarion—Clipperton ore province toward the Guatemala Trench. Isotope data reflected a similar trend.

The presence of buried nodules in Quaternary deposits and their high growth rate were first noted in the Guatemala Trench from samples taken in the MANOP-H map sheet area [21]. Most of the cores were obtained by percussion drill pipe from a depth of 40-50 cm. Nodule diameters exceeded 6 cm. Underlying nodules occurred in carbonate-rich sediments of probably 60-100 ka age. Based on the time of the burial the nodule growth rate was 1200 mm/Ma. Nodule growth rate determinations by U-series methods (^{230}Th , $^{230}\text{Th}/^{232}\text{Th}$, ^{231}Pa) provided of 50-200 mm/Ma values [21], while Mn/Fe ratios yielded a 30 mm/Ma rate for the upper part, and 1050 mm/Ma for the lower part of the nodules [28]. Our data produced a 107-487 mm/Ma mean growth rates for the buried nodules (Table 4).

Isotope data indicate 5-16 mm/Ma nodule growth rates in the MANOP-S map sheet area of the Clarion—Clipperton ore province [31]. Similar values were obtained from other nodules in the north Equatorial zone [27, 34, etc.]. Based on the time of burial, growth rates of buried nodules in the Clarion—Clipperton province (Table 4, Stations 3903-3920) were calculated as 18-87 mm/Ma. Mn/Fe ratios produced a 6.8-20 mm/Ma range. Biostratigraphic and paleomagnetic analyses [13, 14] of surface nodule growth rates in the area produced 16-35 mm/Ma values.

Increased nodule-growth rates in the Guatemala Trench, the far eastern part of the Equatorial belt, resulted from increased biological productivity of the waters and corresponding intensive diagenesis, due to organic matter accumulation and additional introduction of hydrothermal Mn [12, 20].

Comparison of data thus indicates that nodule growth intensity, based on indirect geological information, was 2-5-times greater than established by radiometric means. Growth rates of surface and buried nodules (Table 4) are one order of magnitude (Guatemala Trench), respectively, two orders (Clarion—Clipperton region) lower than sedimentation

rates (Table 1). The figures express mean rates for a given time interval, bracketed by chronostratigraphic horizons. Specific sediment accumulation and nodule growth rates show great variability within such intervals. Buried nodules apparently developed during times of reduced sedimentation rates. Increased deposition rates resulted in reduced nodule growth. This process resulted from global climate changes and, to a greater extent, from local variations in deposition due to sediment erosion by bottom currents.

The relationship between nodule growth rates and sediment accumulation requires a mechanism that promotes it on the mud surface and on the top of semiliquid sediment layers. A number of mechanisms were suggested that would maintain nodules on sediment surfaces: the activity of benthic organism [23, 25, 32, 37, 38, etc.], sediment erosion by bottom currents [3, 18], and the physical characteristics of underlying deposits [35]. We believe that the physical characteristics of the sediments, in combination with bottom current activity [11], are the most likely factors in nodule development.

* * *

1. The data indicate that buried nodules are typical of pelagic sedimentary sequences.
2. The presence of large buried nodules, differences in the structures the composition of the ore matter of surface and buried nodules indicate that during the Pleistocene the nodules had passed through several development stages.
3. Dates, obtained by indirect geological methods on burial conditions, biostratigraphic and paleomagnetic information on nodule growth rates indicate the dominantly Pleistocene age of most diagenetic surface nodules in the northern Equatorial Pacific area.

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RADIOLARIAN FOSSILS AS GUIDES TO THE HISTORY OF SEDIMENTATION OF CHERTY CARBONATES

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UDC 563.14:553(56+57)

An attempt was made to detect trends in the evolution of radiolarians in order to determine the conditions of life in the distant past, which is necessary for an understanding of the temporal and morphological constraints of these unique protozoans on carbonate and silica deposition.

Among the planktonic fauna useful for lithostratigraphy, no single group is known in deposits of all of the Phanerozoic systems. Radiolarians are an exception to this rule: they were present from the Cambrian to the Quaternary. Despite the remarkable diversity of radiolarians in the modern oceans and in ancient sediments, however, the overall pattern of evolution and the ecological restrictions on this group of single-celled organisms are not yet well understood.

The morphology and oryctocoenosis of radiolarians has varied through time and space with the transition from deep to shallow water. But if the depth of deposition of modern siliceous carbonate muds is known [10], and is presumed for the Mesozoic [3], for the Paleozoic it is still very much up in the air [18]. Some investigators believe deposition to have been in the deep oceans [11], others propose shallow water [17, 18, 20].

The hypothesis concerning the predominance of shallow epicontinental seas in the far distant past has recently received new supporting evidence. An analysis of radiolarians can also provide interesting data on this subject.

As the Earth progressed from the depths of time, from the Precambrian to the Quaternary, the relief has increased, the differences between rises and basins has continuously grown larger, mountain ranges have been raised higher, and ocean basins have tended to acquire abyssal depths. The modern hypsometric curve of the Earth's surface is more highly differentiated than in the entire Phanerozoic. All these changes in relief have coincided with the growth of the hydrosphere. Comparison of Paleozoic and Mesozoic–Cenozoic marine basins shows that epicontinental seas, periodically covering vast areas of the continents, have gradually decreased in size through time. At the beginning of the Quaternary they were reduced to seas on the continental margins while the modern oceans were being formed [18, 20].

EVOLUTION OF SEDIMENTATION

The overall geomorphic development of the Earth is not reflected in the evolution of sedimentation.

Chert Deposition. The geological history of chert deposition falls naturally into three major stages: 1) Precambrian and especially Early Cambrian, 2) Cambrian to Jurassic, and 3) Cretaceous to Quaternary. We are concerned here with only the latter two. Comparison of oceanic and geosynclinal silicites has shown that they cannot be considered homologous, and that chert deposition underwent certain changes during the course of the geological history of the Earth [6, 8, 21].

Differences in the chert deposition of the modern oceans and early geosynclines can be explained in three principal ways: the evolution of volcanism, the nature of the basin of sedimentation, and the development of organisms [6, 8, 21–23].

Silicites of the Paleozoic and Mesozoic are either spongiolites, radiolarian jaspers and cherts, or less commonly phthanites (cryptocrystalline siliceous rocks). Diatomites dominated in cherts from the Cenozoic on. This major change in the taxonomic composition of siliceous rock-forming organisms was evidently the result of the overall evolution of the fauna.

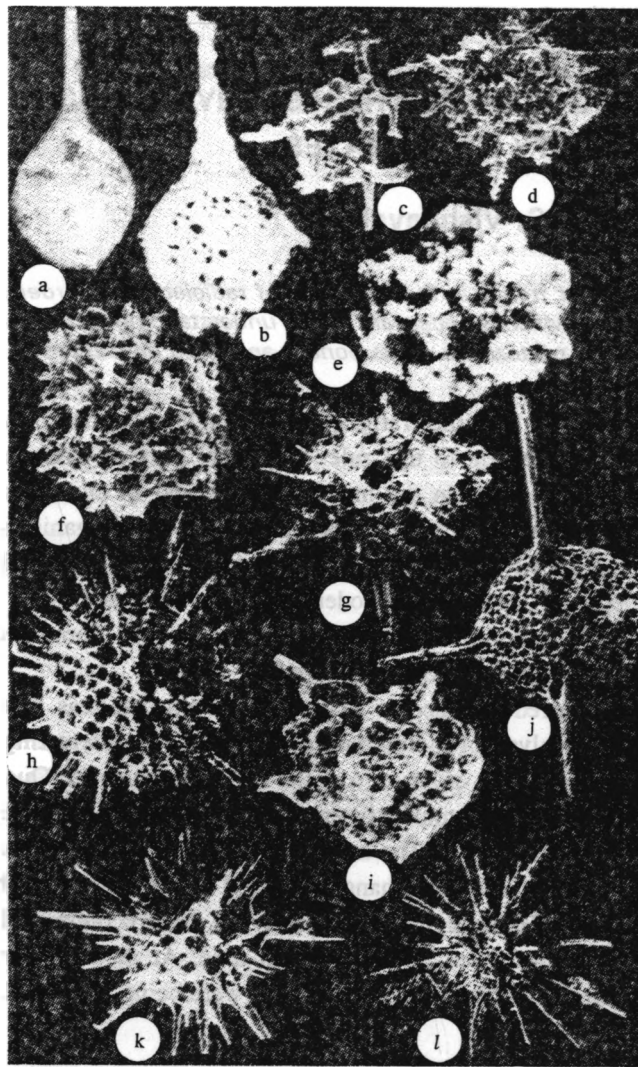


Fig. 1. Radiolarians of the first evolutionary period (Cambrian, Ordovician—Early Devonian). a) Early Cambrian, Atdabanian stage, Kuznets Alatau, Batenev Range: *Lithapium tessiensis* Nazarov, $\times 100$ [13]; b) Early Ordovician, Arenig stage, central Kazakhstan: *Proventocitum procerulum* (Nazarov), $\times 100$ [13]; c-f) Early to Middle Ordovician, Choloi suite, Northern Tien Shan, Karakatty Range, $\times 180$: c, *Ceratoikiscum? acantogulatum* Nazarov (spec. VNIGNI No. 229/8436), d, *Inanigutta eiklaensis* Nazarov (spec. VNIGNI No. 229/8415), e, *Anakrusa? sp* (spec. VNIGNI No. 229/8475), f, *Haplentactinia estonica* Nazarov (spec. VNIGNI No. 229/8413); g-j) Middle Ordovician, Llandeilo stage eastern Kazakhstan: g, *Cessipylorum apertum* (Nazarov), $\times 150$ [13], h, *Oriundogutta ramificans* (Nazarov), $\times 195$ [13], i, *Haplentactinia juncta* Nazarov, $\times 190$ [13], j, *Inanigutta complanata* (Nazarov), $\times 115$ [13]; k, l) Silurian, Wenlock-Ludlow stages, southern Urals: k, *Secuicollacta* sp., $\times 220$ [13], l, *Inanihella tarangulica* Nazarov, Ormiston, $\times 90$ [13].

Deepwater silica deposition in the Mesozoic was somewhat restricted in areal extent, giving way to pelagic carbonates. In modern oceans silica deposition is restricted to east-west belts of high productivity of siliceous organisms. These are determined by climate zones, ocean circulation, and zones of upwelling.

Carbonate Deposition. Carbonate deposition has also experienced significant changes during the Earth's history [20]. Because of the widespread secretion of lime by organisms, chemical precipitation of CaCO_3 gradually halted almost completely during the Phanerozoic, and was replaced by biogenic deposition. As a result, the amount of biomorphic detritus, mud, and other such sediments increased sharply. Although they were formed only in shallow waters during the Paleozoic, by the middle Mesozoic the explosive growth of planktonic foraminifera and algae had supplied carbonate sediments, in the form of globigerina and coccolithophorid oozes, to the abyssal depths of the oceans and major seas.

Comparison of carbonate deposition in ancient and modern geosynclinal basins shows that in both cases it took place mainly in shallow water, on ocean rises or plateaus or in shallow depressions [20].

In principle, the system of shallow-water geosynclinal deposits has analogs in the sediments of the deep oceans. In both cases organogenic sediments predominate.

Slope deposits (carbonate olistostromes and turbidites) are found in both geosynclinal basins and the deep oceans.

Pelagic carbonate sediments of the oceans differ from those of the geosynclines. Cenozoic pelagic carbonate muds are mainly the skeletal remains of planktonic microorganisms, the foraminifera and coccolithophores. They are the chief components of the sedimentary cover on the ocean floor and are distributed over vast areas down to a critical depth, 4.5-4.7 km. Mesozoic Tethys sediments are found that are similar to modern muds, although they tend to occur in the continental areas. Such pelagic carbonate deposits are not present in Paleozoic geosynclines.

Paleozoic deep-water deposits consist of laminated, generally dark-colored micritic and muddy limestones. Such sediments are typical of shallow basins rather than abyssal environments.

The following patterns are characteristic of pelagic geosynclinal basins. During the early and middle Paleozoic siliceous radiolarian muds predominated. Pelagic carbonate sediments were very minor, but often contained a considerable amount of organic matter. Carbonate sediments increased in the late Paleozoic, and in the Mesozoic became the dominant component of the abyssal facies. This distinctive and important change in the composition of pelagic muds can be explained by the massive development and widespread distribution of calcareous plankton and nannoplankton during the Mesozoic, and apparently also to a change in the critical depth of carbonate deposition.

ECOLOGICAL ADAPTATION OF RADIOLARIANS

Organisms are extremely closely related to their environment and living conditions. Thus, as radiolarians shifted from a benthonic (sessile) form to a planktonic one (free-floating or even migrating through different layers of the water), they were freed from an association with certain types of sediments: from predominantly shallow-water siliceous and calcareous deposits in the Paleozoic to relatively deep-water siliceous and calcareous sediments in the Mesozoic and then to deep-water siliceous muds in the modern oceans.

Paleozoic Radiolarians. Radiolarian studies have changed the standard simplistic, nonstratigraphic view of this Protozoan group [2, 13, 25]. The oldest radiolarians were found by Nazarov [12, 13] in siliceous carbonate deposits of the Lower Cambrian Atdabanian stage in the Batenev Range, Kuznets Alatau, and also in Lower Cambrian siliceous rocks in the Azyr-Tal Mountains of Kazakhstan. They are in the shape of hemispheres with pylomes and a long spine, so that they resemble maces. These radiolarians apparently were benthonic; across from the pylomes were pseudopods that fixed the organism to the substrate (Fig. 1a, b) [13, 14]. This hypothesis is consistent with the paleogeography of the sedimentary basin, which indicates that siliceous carbonate sediments accumulated on the floor of the transition zone from an open ocean basin to an enclosed basin [16-18].

Study of the Paleozoic facies and paleogeography indicates that most of the sediments were deposited in shallow marine basins [17, 18]. "Geosynclines are seas filled with islands, the erosion of which supplied the adjacent basins with clastic sediments" [17, p. 244]. Marine sedimentation was widespread during the Paleozoic, both in geosynclinal belts and on shallow marine platforms. In the Early Paleozoic, however, siliceous spicule-algal-radiolarian oozes were predominant, and pelagic carbonate sediments were sparse.

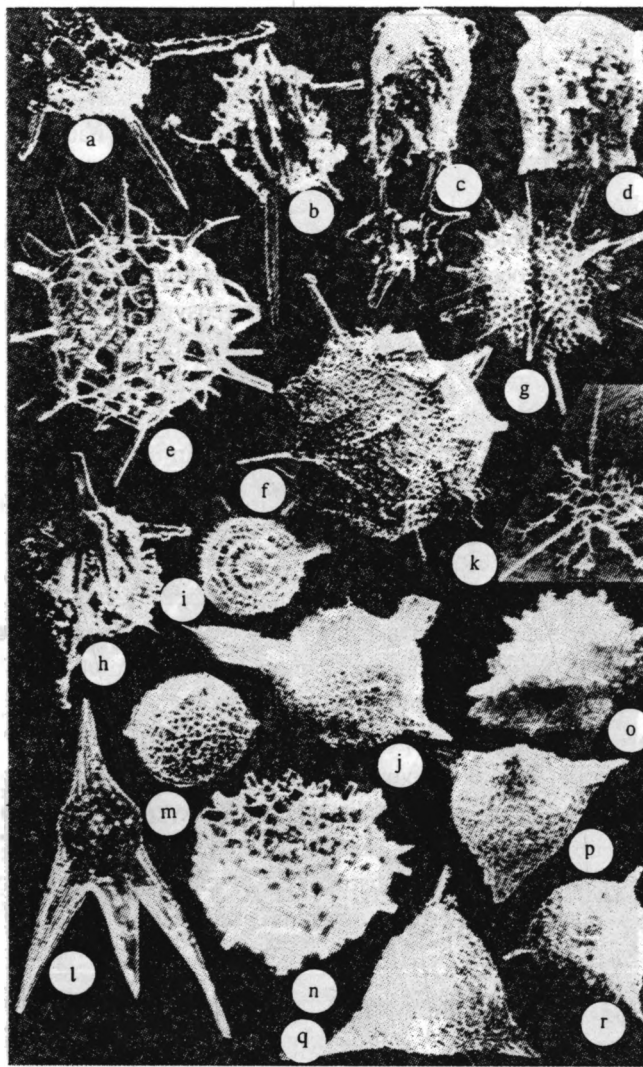


Fig. 2. Radiolarians of the second evolutionary period (Middle Devonian-Middle Carboniferous). a-h) Late Devonian, Frasnian stage (a-d, Polar Urals, e-h, southern Urals): a, *Ceratoikiscum* cf. *bujugum* (spec. VNIGNI No. 229/1806), $\times 80$; b, *Ceratoikiscum* sp. (spec. VNIGNI No. 229/1807), $\times 100$; c, *Holoeciscum* sp. (spec. VNIGNI No. 229/1805), $\times 110$; d, *Caspiaza collarecostulata* (spec. VNIGNI No. 229/3405), $\times 200$; e, *Polyentactinia circumretia*, $\times 250$ [13]; f, *Polyentactinia cosistekensis*, $\times 225$ [13]; g, *Astroentactinis stellata*, $\times 225$ [13]; h, *Ceratoikiscum* sp. (spec. VNIGNI No. 229/1808), $\times 110$; i-l) Late Devonian, Fammenian stage, Pripyat basin, $\times 100$: i, *Pluristratoentactinia conspissata* (spec. VNIGNI No. 229/0521); j, *Tetrentactinia barysphaera* (spec. VNIGNI No. 229/0602); k, *Palaeothalomus* sp. (spec. VNIGNI No. 229/0505); l, Early Carboniferous, Visean stage, France, Monien-Noir [13], *Archocyrtium riedeli*, $\times 250$; m-r) Early Carboniferous, Serpukhovian stage (m,n, northern Caspian; o-r, Tien Shan), $\times 110$: m, *Spogentactinia* sp. (spec. VNIGNI No. 229/0215); n, *Astroentactinia paronae* (spec. VNIGNI No. 229/0205); o, *Caspiaza aculeata* (spec. VNIGNI No. 229/0311); p,q, *Tormentum ruestae* (p, spec. VNIGNI No. 229/0407; q, spec. VNIGNI No. 229/0308); r, *Tormentum sequilateralis* (spec. VNIGNI No. 229/0334).

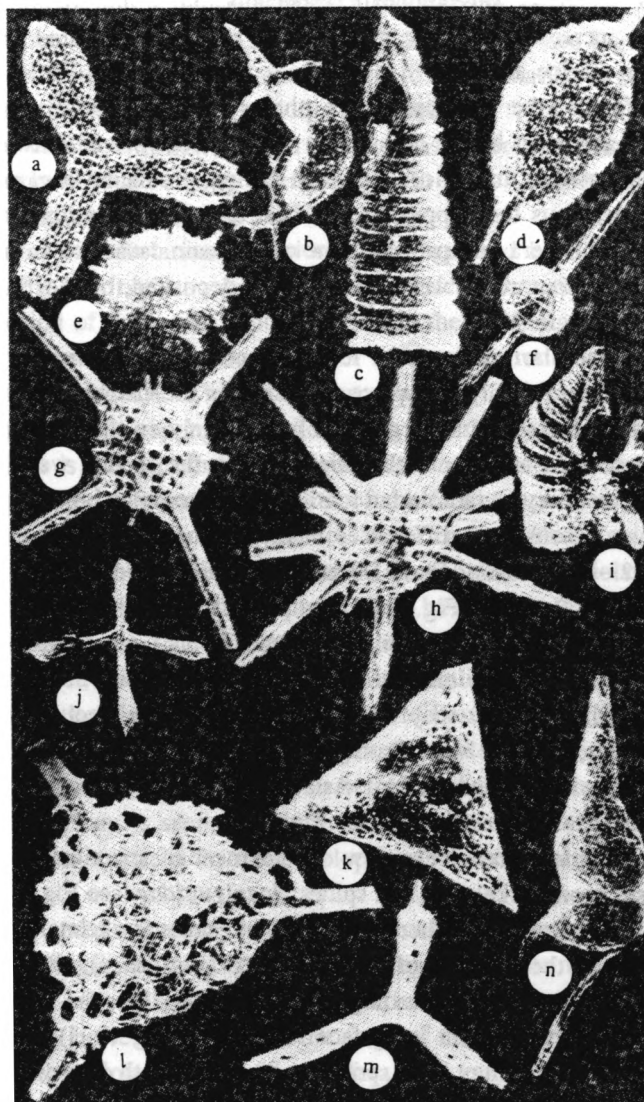


Fig. 3. Radiolarians of the third evolutionary period (Late Carboniferous-Permian). a-c) Late Carboniferous, Gzelian stage, southern Urals: a, *Latentifistula patagilaterala*, $\times 100$ [13]; b, *Haplodiacanthus circinatus*, $\times 95$ [13]; c, *Arrectoalatus cernuus*, $\times 225$ [13]; d-l) Early Permian, Artinskian stage (d, f-l, southern Urals; e, northern Caucasus foothills): d, *Copiellintra* sp. (spec. VNIGNI No. 229/5204), $\times 190$; e, *Copicyntra acilaxa* (spec. VNIGNI No. 229/0337), $\times 100$; f, *Entacinosphaera strangulata* (spec. VNIGNI No. 229/5201), $\times 55$; g, *Entactinia* sp. (spec. VNIGNI No. 229/4207), $\times 140$; h, *Astroentactinia insecta* (spec. VNIGNI No. 229/6908), $\times 130$; i, *Camptoalatus monopterigus*, $\times 150$ [13]; j, *Quinqueremis arundinea*, $\times 25$ [13]; k, *Ruzhencevispongus plumatus* (spec. VNIGNI No. 229/3510), $\times 130$; l, *Tetragregnon scalpratus* (spec. VNIGNI No. 229/4204), $\times 170$; m, n) Late Permian (North America, Delaware Mountains): m, *Follicuculus ventricosus*, $\times 180$ [13]; n, *Latentifistula densa*, $\times 230$ [13].

Silica deposition was the principal element of sedimentation in Early Paleozoic seas. In addition to jasper, related to submarine volcanism, carbonaceous cherts were abundant, especially in the Silurian and Devonian. This particular time interval is also characterized by an increase in organic matter. This is probably related to an increase in carbon dioxide [11, 21], which was concentrated in marine algae.

Later, during the Carboniferous, the development of photosynthesizing land plants apparently led to a further increase in CO₂ in the atmosphere and even in the oceans. As a result, the pH of seawater increased to modern levels (7.5-8.5) [11]. These changes contributed to an explosive growth in lime-producing organisms. At the same time the first true benthonic foraminifera with calcareous tests appeared, the best examples of which were fusulinids. The appearance and increase in calcium secreting foraminifera was probably one of the main causes of an increase in carbonate sediments. Thus, in the Late Paleozoic, while siliceous rocks were being widely deposited, biomorphic carbonates deposits began to increase [8, 20-22].

Under these circumstances, radiolarians arose as follows: spherical shells were formed, which were gradually transformed into free-floating tests. This probably took place at the end of the Cambrian. It is possible, however, that some radiolarians remained attached to the substrate, since organisms lived mainly on the shelf in rather warm marine basins [2, 13]. Moreover, foraminifera that occurred together with radiolarians in the section were solely shallow-water benthonic forms [2]. On the basis of a distinct tendency in radiolarian fossils toward complex but delicate tests, considerable variation in morphology of certain radiolarian taxons, and differences in their association, Nazarov [13] established three distinct periods of development: Cambrian?–Ordovician–Early Devonian (see Fig. 1), Middle Devonian–Middle Carboniferous (Fig. 2), and Late Carboniferous–Late Permian (Fig. 3). These periods are important not only for their stratigraphy, but also for the establishment of stages in the formation of siliceous deposits.

First Period (Ordovician–Early Devonian). Ordovician radiolarians are common in limestones of Kazakhstan, the Urals, the Baltic area, and Spitzbergen; but they are most abundant in siliceous rocks of the northern Tien Shan, Kazakhstan, the Urals, North America, Central Asia, Scotland, Japan, and Kamchatka. Silurian radiolarians are found in carbonate concretions in Canada, the Urals, and Central Asia, but mainly in siliceous shales of France, the northern Caucasus, the southern Urals, Bavaria, and Central Asia (see Fig. 1).

This period was generally a time of relatively slow development of spherical Spumellarians with a thick-walled internal capsule, a porous or coarse, spongy skeleton, and spines: *Inaniguttidae* (see Fig. 1h, j) and *Paleoscenidiidae*, with a spiny skeleton (see Fig. 1c, d). In the Ordovician there was a sort of excess in skeletal variants: the formation of internal spheres that probably took over part of the function of the outer shell (Fig. 1h, i); the raising of the pylomes above the substrate or keeping them away from it by the surrounding spines (see Fig. 1g); the formation of spherical skeletal tests and accessory spines that permitted free movement in surface waters (see Fig. 1h, j); the development of a spiny form, bilateral symmetry, and attempts to form a spongy layer, possibly for use in both deep water layers and in turbulent waters near shore (see Fig. 1c, d, i).

Not all of the variants were successful, however, and by the end of the Silurian almost all of the Early Paleozoic radiolarians had disappeared. Nevertheless, we believe that the radiolarian skeletal variants described above clearly indicate the following: 1) the rapid expansion of highly productive shelf environments as a result of global transgression during the Ordovician; and 2) an apparently considerable lowering of the oxygen minimum in the sea, thus promoting the growth of benthonic forms [11, 27]. Such physical changes in the environment of any organism, including radiolarians, must produce an alteration of the skeleton that is a better adaptation to the new ecological conditions.

Second Period (Middle Devonian–Middle Carboniferous). The deepening and expansion of marine basins in the second half of the Paleozoic had an effect on the development of radiolarians (see Fig. 2).

The best known well preserved Devonian radiolarians are in carbonate concretions and limestone lenses in the Urals, Kazakhstan, Australia, Canada, North America (Ohio), and Kolyma. They are less common in cherts in the Urals, Central Asia, Kazakhstan, Japan, and North America (Alaska). Carboniferous radiolarians are known in phosphatic concretions in Turkey and southern France, in limestones in North America (Oklahoma), Kamchatka, the southern Urals, the Caspian foothills, and Tien Shan, and in cherts in North America (Alaska, California), Central Asia, Japan, China, Kazakhstan, and the Kamchatka-Chukhotsk region.

The domanikoid facies of the Volga-Urals district and the Timano-Pechora province, the Pripyat depression of the Russian platform, and the Appalachian foredeep of the North American platform have especially abundant and varied radiolarians of this period; they were deposited under conditions of highly undercompensated subsidence. And despite the suggestion of Yulovich [24] that ashfall led to massive loss of radiolarian plankton at this time, resulting in their remains

becoming abundant in carbon-rich beds, we propose instead that these radiolarians had a stable symbiosis with algae. This view is supported by symbiotic relationships between modern radiolarians and algae [14]. Apparently many Devonian radiolarians were benthonic, and many of these were attached to algal surfaces (see Fig. 2a-d, h, i, l, m) [2, 13].

At the same time strong indicators of paleocurrents leading to the formation of contourites in domanikoid and other basins [9] suggest that radiolarians had to transform rapidly to a planktonic form, lightening their tests and modifying the structure, and thus become more common in new, deeper water areas far from shore. In this regard it is significant that true sessile benthonic forms have been found among pilitonemids in Early Carboniferous clastic reefal limestones in the Karachaganak massif, northern Caspian foothills (see Fig. 2m) [1].

Overall, the second period of radiolarian evolution is distinguished by a transition to spherical and bilaterally symmetrical skeletons (see Fig. 2). The rapid evolution of the spherical *Entactiniidae*, which formed a number of new taxons in a geologically short period of time, was expressed first in the lightening of the internal capsule: it became less massive in 4-, 6-, and multirayed spicule forms. On the other hand, alteration of the Entactinids was manifested in the main spines of the skeleton: massive, cylindrical, primary spines (see Fig. 2e, f) became delicate triangular ones (see Fig. 2g, j, m, n). But the emergence of light, spongy skeletons was especially typical of the entactinids (see Fig. 2j, k). Moreover, an increase in variations is found in the Late Devonian *Palaeoscenidiidae* (see Fig. 2h).

This same segment of geologic time saw the advent of the first stauraxonic polycystins (see Fig. 2p-r) and the origin of *Albaillellaria*, apparently a prototype of the Mesozoic and Cenozoic *Nassellaria*.

Third Period (Late Carboniferous–Early Permian). Carboniferous radiolarians have been described chiefly from cherts in North America, Japan, and the southern Urals. They have a high proportion of bilaterally symmetrical and stauraxonic forms. In addition, during the Late Carboniferous a number of bilaterally symmetrical forms developed a clearly segmented skeleton with apical, central, and basal parts (see Fig. 3b, c).

According to the biological laws established in 1851 by Milton-Edwards [14], the evolution of an organism is always accompanied by differentiation of the skeleton and organs. This developmental trend is seen in originally identical organisms gradually becoming different from one another in both form and function. That is, on the one hand a tendency toward specialization of individual organs is seen, and on the other, complexity and polymerization of the organism as a whole. Definite steps along this line are seen in the development of spherical, and especially stauraxonic, Paleozoic radiolarians.

Later, in 1936, Dogel' [7] noted a tendency toward a decrease (oligomerization) in the number of unique homologous structures during organic evolution. It is thus thought that oligomerization is a second step in the evolution only of multicelled animals, whereas single-celled organisms exhibit the initial stage, polymerization [7, 14]. Many radiolarians and foraminifera proceed to the end, however, and reach the stage of oligomerization.

It may be assumed that both radiolarians and foraminifera have very progressive natures with stable skeletons and a small number of skeletal elements. The especially ideal spongy skeletons of Paleozoic radiolarians, constructed out of very fine interwoven strands with a minimum of material and expenditure of effort, are very important for planktonic forms. It is possible that the submergence of basins at the end of the Paleozoic, i.e. the appearance of broad, deep waters, initiating a gradual approach to modern abyssal conditions, led to the origin of oligomerization in radiolarian skeletons, and subsequently in planktonic foraminifera as well. The preferential occurrence of radiolarians in deep-water sediments at the end of the Paleozoic and during the Mesozoic and Cenozoic supports this idea.

Permian radiolarians are widely distributed in cherts of Japan and China, but many are also found in detrital limestones of the southern Urals, the Caspian foothills, and North America (Texas).

Permian time was very unusual and is distinguished by strong development of the stauraxonic radiolarians *Latentifistulidae*, *Tormentidae*, and *Ruzhencevispongidae* (see Fig. 3a, j, k, m), somewhat slower evolution of spherical spumellarians (see Fig. 3d-h, l), and the irregular but very curious development of bilaterally symmetrical forms with distinctly segmented tests (see Fig. 3b, c, i, n). Differentiation of spongy texture, formation of an internal capsule, considerable increase in skeleton size, and extraordinary variability in the external shape of the test are seen in the morphological alteration of spherical radiolarians (Fig. 3).

Although spherical radiolarians were still quite numerous during this time interval, the characteristic feature of the period is the widespread distribution of stauraxonic forms. The chief feature of the stauraxonic radiolarians that distinguishes them from all others before and since is the unique form of the test: from subtriangular to spindle-shaped or

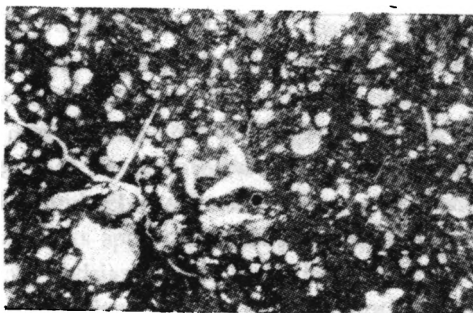


Fig. 4. Cherty limestone. Radiolarians make up 30% of the total volume of the rock. Mediterranean foldbelt. Lesser Caucasus, Nagorno-Karabakh, village of Yanshakh, Maidanchai River basin. Middle Jurassic. Slide 064-56, $\times 25$.

bladed with three, five, or more blades, but united by a single plan for the structure of the internal capsule. It is a hollow sphere with hollow rays extending out of it. This skeletal form apparently arose in order to provide a certain advantage for planktonic organisms and help them achieve a wide geographic distribution.

Stauraxonic radiolarians, which appeared suddenly in the Early Carboniferous and disappeared just as suddenly at the end of the Permian, unfortunately left no successors in the Mesozoic. The exceptional variability among radiolarians at the close of the Paleozoic, especially during the Artinskian stage in the Early Permian, may have been related to the mild climatic conditions. The variety of radiolarians was sharply curtailed at the end of the Permian, however, and at the end of Late Permian time almost all of the known forms of Paleozoic radiolarians had vanished.

Mesozoic Radiolarians. Unfortunately, as yet no true link has been established between Mesozoic–Cenozoic and Paleozoic radiolarian taxons, which experienced almost total extinction at the end of the Permian. Moreover, shallow-water environments of sedimentation in the Triassic and Jurassic were quite similar to those in the Paleozoic, both in the present ocean basins and on the continental blocks. Subsidence of the ocean and uplift of the continents that began during the Mesozoic brought about a restriction of the seas, however. At the same time, apparently epicontinental seas contracted and shorelines were stabilized [18].

In turn, orogenic activity and a general uplift of the continents all over the globe during the Mesozoic reduced carbonate sedimentation to a minimum.

Thus, the beginning of the Triassic, marked by shallow-water carbonate sedimentation in the Tethys, gradually changed over to coal deposition (Romania, Italy, Yugoslavia, and elsewhere). Triassic radiolarians have been found both in carbonate rocks of central Europe and in clastic limestones of Central Asia and thin micritic limestones of the Far East and North America. They are all spongy rings and discs, multirayed spheres, and less commonly parachutes with high apices. Siliceous carbonate deposits became more abundant in Tethys in the Middle Triassic. Light-gray limestones are predominantly micritic, and often associated with thin lenses of chert and microconcretions of black chert. Along with chert, the limestones contain interbeds of green pelitic tuffs. Radiolarians at this time were primarily spheres with spiral twists of two to three, five to six, or even a single spicule in the shape of a braid or corkscrew ([28], p. 162; see *Bogdanella trentana* Kovar-Yurkovsk). These spicules apparently enabled Triassic radiolarians to float freely in the water or to change from benthonic to planktonic and back by means of a rotating motion of their axopods, and were quite effective in clear, warm, shallow water.

Only at the end of the Middle Triassic did carbonate deposition begin to increase again; it reached a maximum at the end of the Cretaceous. This latter was probably due to a new, explosive growth of limestone-secreting organisms, foraminifera and coccolithophores. Like the fusulinid bloom during Late Carboniferous–Early Permian time, after a certain lull the rapid build-up was repeated in the Cretaceous. A global and, geologically speaking, instantaneous expansion of calcareous micro- and nannoplankton, just as in the Paleozoic, brought about massive deposition of carbonate sediment. Foraminifera were predominantly planktonic at this time.

Modern basins show that the presence of radiolarians in carbonate sediments is always an indicator of a connection with the open sea. This connection first appeared most clearly in the Mesozoic.



Fig. 5. Detrital (spongy, coralline, radiolarian) limestone. Lesser Caucasus, Little Chaikend Mountain. Late Jurassic. Slide 115-5, $\times 25$.

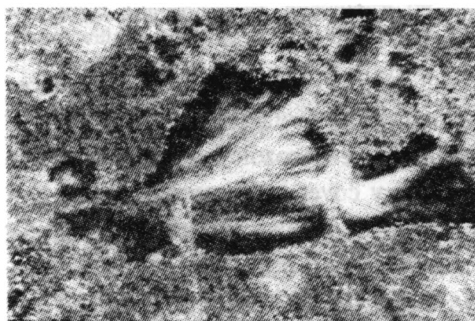


Fig. 6. Spongy, algal, radiolarian limestone. Lesser Caucasus, Lysogor Pass. Late Jurassic. Slide 114-2, $\times 25$.

If Mesozoic sediments are examined from this point of view, then it is clear that radiolarian ooze tends to occur in the relatively deep parts of the ocean basins and carbonates in relatively shallow parts and along the sea coasts. Recently, as a result of detailed microscopic study of numerous Mesozoic limestones of the Caucasus, the Carpathians, Central Asia, and the Far East, their primary biogenic nature has been determined [5, 23, 26, 29].

It has thus been established that, along with coccolith and foraminiferal muds, Mesozoic and Cenozoic slope sediments contain mixed radiolarian-foraminiferal and spicule-radiolarian-coccolith siliceous carbonate muds. Radiolarians in carbonate sedimentary rocks in the Mediterranean foldbelt sometimes make up 25-50% of the total (Fig. 4). Examples include light-gray Cenomanian–Campanian limestones of the Tuapsin region, Turonian-Maastrichtian of the Kutaiss region (Surami and other sections) and western Abkhazia, Tithonian–Neocomian siliceous limestones of the Susuzlykh and Sarybabin zones of the Lesser Caucasus, Feodosia region of Crimea, and others [3-5].

Radiolarians in Middle Mesozoic carbonate rocks are restricted mainly to micrites, although they are sometimes encountered among clastic porous algal limestones (Fig. 5, 6). Early Jurassic radiolarians have been found in the Alps, the Caucasus, and in the Pacific rim countries (Japan, eastern Russia, western USA). Middle and Late Jurassic radiolarians are well known in fine, micritic limestones of Italy, Yugoslavia, France, Spain, and Bulgaria, as well as in the Crimea, the Caucasus, the Carpathians, and the Hissar Mountains. They are commonly found in the Pacific province: the Koryaks, Kamchatka, California, Japan. Early Jurassic radiolarians include both high conical forms of the primitive nassellines *Canoptum* and *Dictyomitrella* and numerous low conical forms, often with well formed pseudoforamen, of *Bagotum*, *Canutus*, and *Droltus* [29]. An extra opening in the shell walls probably made planktonic wandering possible for the radiolarians (from the Greek "plankto," to wander) by adhering to algae or other objects with the aid of pedicles or suckers, which appeared out of the so-called windows or openings in the form of pseudopoda. Thus, this opening permitted these Early Jurassic radiolarians to lead both a sessile benthonic life and also a planktonic life. The pseudopoda were possibly inherited by the Jurassic nasselline radiolarians from Paleozoic bilaterally symmetrical albaillellarians. The

highest beds with a large amount of radiolarians adapted to both benthonic and planktonic environments is Bajocian in age. At this particular time the first known truly planktonic foraminifera appeared. It is therefore natural to assume that both radiolarians and planktonic foraminifera first conquered the open ocean in the Middle Jurassic, contributing to oceanic carbonate and silica sedimentation. Late Jurassic radiolarians are characteristically all planktonic [3].

Cretaceous radiolarians are found everywhere in carbonate sediments of the world's oceans (drill holes 367, 466, 534, and others), in foraminiferal limestones of Germany, Romania, the USA, China, Japan, the Caucasus, western Siberia, and the Far East, and also in cherts and jaspers.

To the extent that the environment changed from shallow to deeper water, changes are found in the morphological features of the radiolarian tests. Typical of the Cretaceous are towerlike nassellines and spherical or discoidal spumellarians in association with planktonic and, rarely, benthonic foraminifera. Smooth and spongy skeletons dominated by two or three morphotypes are typical of pelagic beds closer to shore. It is well known that a high percentage of multicyrtoid nassellines in fine-grained Mesozoic and Cenozoic carbonate muds indicates great and often contrasting depths of the basin. Cretaceous carbonate sediments of the Central Atlantic, the Hess Plateau, the Northwest Pacific, and elsewhere are examples [26, 29]. Spiny radiolarian forms are particularly characteristic of moderately deep carbonate beds. Radiolarians with split extensions, which are dominant in an oryctocoenosis of discoidal forms, may indicate near-shore carbonate deposition. Many Callovian–Kimmeridgian limestones of Central Asia and the Greater Caucasus are similar [5], as are Lower Permian limestones of the southern Urals [12].

Cenozoic Radiolarians. As a result of crustal extension and subsidence of vast areas of the American, European, African, and Australian continents at the end of the Cretaceous and beginning of the Paleogene the modern oceans began to take shape. Deepening of the ocean basins in the Cenozoic led to the development of a truly oceanic sedimentation. This event was imprinted on the oceanic sedimentary cover and preserved traces of the gradual development of the sediments, from very shallow, typically lacustrine-marine, to regular oceanic [18]. The Cenozoic is marked on the whole by sharp changes in sedimentation. It is characterized by the restriction of chert deposition to the continental margins and the widespread occurrence of abyssal pelagic carbonate muds.

Approximately in the middle of the Cretaceous, at almost the same time as the deposition of deep-water oceanic sediments, siliceous sedimentation changed from "geosynclinal" to platform. This was apparently in response to a bloom of diatomaceous algae in epicontinental seas. Biogenic silica deposition continued in the oceans at this time. This continued to increase throughout the Cenozoic and culminated in the formation of modern siliceous diatomaceous-radiolarian oozes in the oceans [21, 22].

Diatomaceous algae, various radiolarians, and silicoflagellates are the principal concentrators of silica. Three quarters of the silica in the oceans was derived from the shells of these organisms in three east–west belts of sedimentation: southern, corresponding to the southern humid zone and composed mainly of diatom shells; equatorial, extending along the humid-tropical zone and including both diatomaceous algae and radiolarians; and northern, a moderately humid zone of predominantly siliceous diatoms. The position of these belts is closely controlled by the areas of maximum plankton productivity. In addition to these belts, there are certain other spots with intensive silica deposition located in areas of upwelling. These are found mainly along the west coast of Africa, along the coast of Chile and Peru, and even at the latitude of California [22].

Modern carbonate sedimentation in the oceans is almost entirely biogenic, with the major contributors being planktonic foraminifera and coccolithic algae.

In analyzing the distribution of modern sediments, it can be seen that foraminiferal and mixed radiolarian-foraminiferal oozes are restricted to relatively shallow areas of the ocean, while purely radiolarian oozes are present in deeper waters. This is partly related to the depth limit of carbonate solution, but also to the restriction of the high biological productivity of radiolarians to east-west belts. Radiolarians are practically unknown in Cenozoic shallow-water limestones.

Cenozoic radiolarians include nassellines, spumellarians, deep-water phaeodarians. They occur mainly as spongy, discoidal, ellipsoidal, and spherical forms, and also impoverished assemblages; in shallow-water neritic regions, they are dominated by one or two types. Such restricted associations generally indicate shallow water and nearby land areas. On the other hand, morphological variability and large amounts of nasselline tests suggest considerable water depth and the presence of upwelling. These patterns were established from modern studies [10] and are well displayed in Mesozoic radiolarian assemblages.

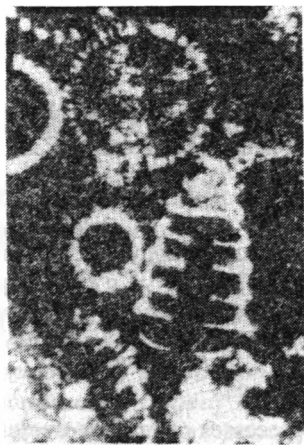


Fig. 7

Fig. 7. Managaniferous radiolarian jasper. Internal voids of radiolarians and groundmass composed of manganese (black: psilomelane, pyrolusite needles within radiolarians). Cretaceous, Lesser Caucasus, village of Kaplykend. Slide 047-3, $\times 50$.

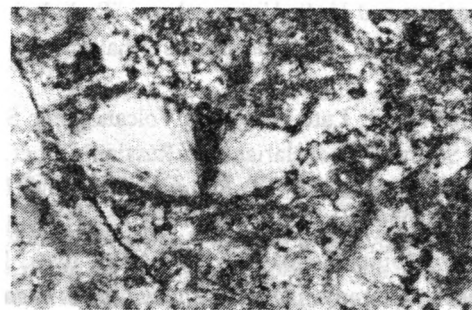


Fig. 8

Fig. 8. Siliceous limestone. Fragments of chalcedony surrounded by organic matter. Jurassic, Lesser Caucasus, village of Sarybaba. Slide 111-9c, $\times 25$.

One other group of modern radiolarians is known: the acantharians. These radiolarians have a celestite skeleton and are predominantly neritic. A considerable number of modern acantharians live in the Mediterranean Sea, but there are also oceanic populations. Fossil acantharians are quite rare because their skeletons are so easily dissolved. Only one large deposit of their tests, composed of strontium sulphate, is known, in Eocene sands of Karakum. Fersman [19] proposed a radiolarian origin for celestite interbeds in Permian limestones of the Volga—Urals region.

Thus, the presence of radiolarians in carbonate sediments will always be taken to indicate a connection with open marine waters. And the presence of large numbers of radiolarians in fine-grained carbonate sediments, lacking turbidite textures, will primarily suggest a low rate of sedimentation and often the presence of nearby ore mineralization (manganese, phosphorous) (Fig. 7). Examples include Devonian and Carboniferous limestones of Pai-Khoya and the polar Urals [15], radiolarian-foraminiferal carbonate oozes (holes 465, 466) in the Pacific Ocean, and others. Relatively deep-water radiolarian communities in the ocean-continent transition zone are often associated with carbonate rocks with a high content of organic matter (Fig. 8), which outlines siliceous fossil fragments and is concentrated in the porous radiolarian skeletons or fragments.

* * *

The rock-forming role of radiolarians confirms the direct connection between organism and sediment. Radiolarians are the only faunal group that has actively contributed to sediment throughout the entire Phanerozoic, forming thick sequences of jasper and phthanites or domanikites in the Paleozoic, and jasper and subsequently chert or chalcedony in the Mesozoic and Cenozoic.

The close association of Paleozoic radiolarians and algae (domanikites) and benthonic foraminifera is followed by a thanatocoenosis of Mesozoic—Cenozoic radiolarians and diatoms, coccolithophores, and planktonic foraminifera. The bios may include both planktonic and benthonic (or transitional) forms of radiolarians. The suggestion that ancient Paleozoic cherts are of relatively shallow-water origin is therefore just as valid as the view that there are deep-water jaspers and cherts.

Thus the appearance of benthonic or transitional radiolarians in the early Mesozoic and Paleozoic (forerunners of the modern siliceous or celestitic organisms) sheds light on global changes in silica and carbonate deposition and, most significantly, on the bathymetry of ancient oceans.

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THE PROBLEM OF THE ORE CONTENT OF TIMAN

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On the basis of generalization of new data, this article considers the problem of primary gold and diamond content of the Baikalian basement complex of the Timan uplift of the Russian platform, and also the conditions of distribution and formation of polymineral placers in the Paleozoic platform mantle. A conclusion is drawn about the probable existence of an extended ore zone in the region, which is promising for discovery of primary gold and diamond deposits, as well as paleoplacers of gold, diamonds, and rare-metal and rare-earth minerals in Devonian conglomerates and gritstones.

Data obtained as a result of research by the authors of the present article and generalization of materials from previous works of the Central Scientific Research Geological-Survey Institute (TsNIGRI), the All-Union Scientific Research Geological Institute, the Komi Branch of the Russian Academy of Sciences, the Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry of the Russian Academy of Sciences (IGEM RAN), and the Ukhta and Timan expeditions of the "Polyarnouralgeologiya" and Arkhangel'skgeologiya" enterprises [1, 7, 11-13, 15-20, 22] enabled us to come to the conclusion that the Timan chain of hills is an extended ore-bearing zone in which, besides the known deposits of Devonian bauxites and large titanium paleoplacers, gold-ore deposits can be found in carbonaceous shales, and placers of gold and diamonds in conglomerates, similar to those already revealed in Central Timan (the Ichet-Yu area).

The chain of hills is located in the northeastern part of the Russian platform and is sharply distinguished tectonically from its central parts by the younger, Baikalian age of the basement complex. The regional structure under consideration is one of the branches of the Urals' Baikalides [5, 6, 8, 21], comprising a chain of uplifts stretching almost 1000 km in a north-northwesterly direction from Polyudov Kamen in the Northern Urals to the Kanin Peninsula, with a width reaching 120 km. Timan is separated from the adjacent Russian platform, with its Lower-Precambrian crystalline basement complex, by a marginal junction, which, according to indirect geological data, is a large overthrust; and from the Paleozoic-Mesozoic Pechora syncline, by a system of deep flexures and faults. The uplifts of Southern, Central, and Northern Timan are dome-block platform structures of varying size, in the cores of which the Baikalian folded basement complex is exposed; and on the limbs, Devonian terrigenous formations that are the lower horizons of the epi-Baikalian craton's slightly dislocated mantle.

It is precisely these structures that are of prospecting interest, since manifestations of gold-ore mineralization are confined to rocks of the Baikalian basement complex, and lower horizons of the platform mantle are connected with placer occurrences and the first established gold placers with accompanying diamonds and rare-metal and rare-earth minerals in conglomerates and gritstones.

STRUCTURE AND ORE CONTENT OF THE BASEMENT COMPLEX

The Baikalian basement complex is represented to various degrees by dislocated strata of the Upper Precambrian (Middle and Upper Riphean, and, possibly, Vendian). On the whole, the composition of the strata is homogeneous: carbon-containing phyllite-like shales and aleuroshales are predominant, to which quartzites, quartzite sandstones, dolomites, and limestones with bioherm stromatoliths are subordinate. The rocks have experienced regional metamorphism (zonal, in places), cleavage, and, in zones of rupture dislocations, crushing and cataclasis. The total thickness of the cross section has been estimated differently by various researchers, but is probably no less than 10-12 km.

In the structure of the Baikalian basement complex of Timan, two zones with a northwesterly trend are distinguished, separated by a large rupture dislocation [6, 9] (Fig. 1).

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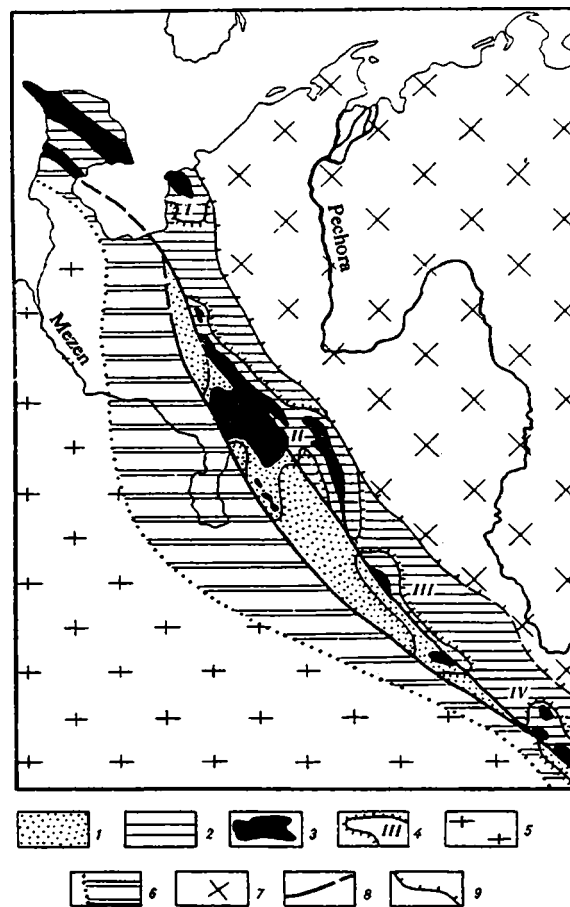


Fig. 1. Diagram of tectonic regionalization of Timan and adjacent territories (according to [6], with changes). 1-4) Timan chain of hills (in a platform structure of megaramparts): 1) outer zone of gently sloping deformations of the Baikalian basement complex, 2) inner zone of linear folding of the Baikalian basement complex, 3) outcrops of the Baikalian basement complex to the ground surface, 4) dome-block uplifts of the epiBaikalian craton (I – Northern Timan, II – Central Timan, III – Southern Timan, IV – Polyudov Kamen, northwestern part); 5-6) Russian platform with Early-Precambrian crystalline basement complex: 5) elevated in the Riphean, 6) submerged in the Riphean (zone of pericratonal subsidences); 7) Pechora syncline with Baikalian basement complex; 8) large rupture dislocations; 9) eastern boundary of the inner zone of the Baikalian basement complex of Timan.

The southwestern (outer) zone was preserved in the contemporary structure mainly within the bounds of the Central and, partly, the Southern uplift of Timan. It is characterized by a relatively elevated position and the predominant occurrence of rock of the lower part of the cross section, represented in Central Timan by the Chelas series of the Middle Riphean (quartzite sandstones, shales), with apparent thickness of no more than 3.3 km. An important feature of the zone is relatively slight dislocations: the strata are deformed into gently sloping (7-10°), wide (15-20 km), in places flat-topped, folds.

The northeastern (inner) zone is traced along the whole of Timan. It is characterized by a submerged position and, accordingly, by a large thickness of strata and linear folding with a northwesterly trend, with clearly manifested vergence of the folds and an upthrust–overthrust structure of longitudinal faults, with uplifted eastern limbs. The latter indicates displacement of the masses during folding in the direction of the more rigid outer zone and the Russian platform located beyond it with a crystalline basement complex – a phenomenon characteristic of many geosynclinal folded structures. In

the inner zone, strata of the middle and upper parts of the Precambrian cross section are developed: the Bystra (3500 m), Kisloruchi (1200 m), Vym and Lunvoga (~6000 m) series, which are experiencing overall subsidence to the east-northeast. The inner zone's tectonic mobility is also confirmed by the presence in schistose strata — aspid and flyschoid formations in cross sections of the Barma series in Northern Timan (Rumyanichnaya suite), and of the Vym and Lunvoga series in Central Timan — of horizons and interlayers of turbidites, which indicate a relatively deep-water situation of the continental slope, at times with uncompensated sedimentation. The inner zone is also characterized by locally manifested intensification of the regional metamorphism of rocks to the epidote-amphibolite facies [6, 9], which is not noted in the outer zone. Thus, in Northern Timan, in the field of development of phyllitelike carbonaceous shales and metaaleurolites of the Barma series, one of the boreholes (No. 112) revealed garnet-staurolite mica schists within the bounds of a local gravitational anomaly. Products of their destruction, in the form of large (up to 15-25 mm), semirounded crystals of staurolite, partially replaced by kaolinite, and round grains (up to 2-3 mm) of almandine enriched with manganese, are strewn through the filler of Devonian quartz-pebble conglomerates along the Malaya Chernaya, Velikaya, and other rivers [14]. We found analogous concentrations of large fragments of staurolite and garnet in Devonian conglomerates in Central Timan along the Umbe River, where garnet-staurolite mica schists are thus far unknown.

In connection with the existence of outer and inner zones, it is relevant to note the confinement of carbonate strata with abundant stromatoliths to the structural junction separating the zones (a deep paleoflexure). This allows us to suppose that the carbonate strata are formations of the barrier-reef type (the Pav'yug suite of the Upper Riphean in Central Timan, etc.) [6] initially developed on the raised limb of the flexure, which probably did not accumulate in the submerged inner zone.

Baikalian folding, as a result of which the basement complex of the Timan platform uplift was formed, most likely occurred in the second half of the Vendian and culminated in the injection of intrusions: in Northern Timan there are small masses of gabbros, granites with molybdenum mineralization, and syenites. According to geophysical data, there are supposed to be undisclosed stocks in Chetlaskii Kamen of Central Timan.

The composition and structure of Upper-Precambrian strata of the Baikalian basement complex of Timan, the order of their thicknesses, the presence of final folding accompanied by cleavage and regional (though slight) metamorphism and also by intrusion of orogenic granites, and molybdenum mineralization are signs that allow us, with sufficient basis, to classify the paleotrough of Timan (its inner zone in particular) as a geosynclinal structure. The similarity of Timan to folded systems of the Enisei chain of hills, the Patom highland, and the Yudoma-Maya trough in the borders of the ancient Siberian craton indicates the Timan belongs to miogeosynclinal folded systems with the gold mineralization in "black-shale" formations that is characteristic of them.

Manifestations of gold mineralization have been established in carbonaceous shales and metaaleurolites of the Baikalian basement complex in Central and Northern Timan and are confined primarily to the inner, linear-folded zone. They are classified as gold-sulfide-quartz and gold-sulfide ore formations, and, to a lesser degree, as a low-sulfide gold-quartz formation.

It is noteworthy that many of the manifestations that have been studied, which belong to the first two formations, being localized in zones of crushing, schist formation, and cataclasis along rupture dislocations, gravitate to intervals of the cross section of terrigenous-shale strata with increased (up to 3-6%) content of C_{org} in shales and metaaleurolites, which are frequently characterized, moreover, by increased (up to 0.08 g/ton) background gold content. This allows us to think that originally sedimentogenic gold dispersed in the rocks may have taken part in formations of zones of superimposed quartz-sulfide vein-disseminated mineralization by means of its remobilization and redeposition under the action of hydrothermal-metasomatic processes. Detailed study of typical mineralized zones in the northern part of Tsil'menskii Kamen (Central Timan), which we conducted together with D. A. Dorofeev and I. Z. Isakovich (TsNIGRI) enabled us to distinguish two different-aged mineral associations, establish typomorphic features of the main ore mineral (pyrite), and reveal a close correlation between fine segregations of high-grade native gold and late fine-grained pyrite, which is accompanied by marcasite, chalcopyrite, pyrrhotine, sphalerite, galenite, a mineral from the group of cobalt and nickel arsenides, quartz, carbonate, and, less often, gray copper ores.

The features of ore mineralization that were revealed make it similar to that in carbonaceous terrigenous-shale strata of the Riphean in the Lena gold-bearing region of the Patom highland. This is expressed in similarity of the outward appearance, textures, and structures of mineralized rocks, their belonging to a low stage of regional metamorphism, the

simple set of ore minerals, the same sequence of their crystallization, the presence of relict ilmenite, leucoxene, and rutile, which are characteristic of ore-containing shales, and in the connection of gold with the late association of neogenic minerals.

An important feature of Timan is that rocks of the Baikalian basement complex, including mineralized zones, experienced intensive chemical weathering of an allitic and siallitic profile at the end of the Early and in the Middle Devonian (which is established by the presence of numerous fragments of crusts, areal, as well as linear). This enables us to suggest the existence above zones of gold-sulfide mineralization, in areas where the erosion cut was not so deep, of gold-bearing oxidation zones, including "iron" caps.

As in the Lena region, part of the gold in Timan is connected with quartz veins and zones of veining. Two varieties of it should be classified as ore quartz: milky-white and gray opaque, the presence of metal in which is established by assaying and confirmed by mineralogical analysis. In areas where Devonian conglomerates are enriched with quartz pebbles of these varieties (as on the southwestern flank of the Ichet-Yu gold field), the content of very-fine and fine placer gold rises noticeably in them. Semitransparent spotty-block quartz is also probably gold-bearing. We were unable to observe it in its original position in the form of veins and veinlets, but in Tsil'menskii Kamen, in part of the conglomerates, its content in the composition of pebbles correlates directly with the content of placer gold.

The slight degree of study of gold-ore mineralization in carbonaceous shales of the Baikalian basement complex does not permit us to judge the scale of mineralization or its areal extent. However, according to indirect data, it is fairly widely manifested in Northern and, mainly, Central Timan (Southern Timan remains poorly studied). This is confirmed by finds of gold in contemporary alluvium of many rivers and creeks, and also by the discovery of the first placer objects in them. On the whole, gold in the alluvium is very fine (0.1-0.25 mm), high-grade (fineness 930-980), and similar to metal from zones of quartz-sulfide and sulfide mineralization in carbonaceous shales of the basement complex and, partly, from quartz veins. But, in areas of widespread development of the quartz-vein type of mineralization (the Kykvozh River north of Vym ridge, etc.), according to V. A. Dudar's data (Ukhta Geological-Survey Expedition), medium and coarse gold (up to 2-3 mm) is predominant in contemporary alluvium.

The role of carbonate rocks of the Upper Riphean for localization of gold mineralization has not been established. However, their position near the permeable zone of large-scale, long-lived dislocation is favorable for manifestations of hydrothermal-metasomatic processes and indicates the possibility of discovery of gold mineralization similar to that of the Carlin deposit (USA) and the recently discovered Tas-Yuryakh object in the Yudoma-Maya trough.

The age of gold-ore mineralization — the end of the Late Precambrian (second half of the Vendian?) — coincides with the conclusion of formation of the folded structure of Timan and the penetration of minor intrusions of gabbros, granites, and syenites mentioned above. The mineralization is confined to rupture dislocations complicating and, in places, cutting across the folded structure. At the same time, the mineralization is known to be older than dikes of alkali-ultrabasic and manifestations of alkaline metasomatism with radiological age, according to N. A. Dovzhikov's data, of 570-590 million years, which cut across gold-bearing zones in Central Timan and are connected with the first epoch of riftogenesis of the neogenic epiBaikalian craton.

The data set forth indicate the prospects of Timan for discovery of gold-ore deposits within the bounds of ledges of the Baikalian basement complex, mainly in the inner structural zone. In type and age, they may prove to be similar to deposits in carbonaceous shales in the Lena gold-bearing region, the Sayan region, the Enisei chain of hills, and also, possibly, to objects in carbonate strata in foreign countries. Besides these, discovery of deposits of the gold-bearing oxidation-zone type is likely in Timan.

STRUCTURE AND ORE CONTENT OF THE PLATFORM MANTLE

The platform mantle is represented by fragmentarily preserved Ordovician (?), Silurian, Lower- and Middle-Devonian, and more widely developed Upper-Devonian, Carboniferous, and Lower-Permian deposits. In contrast to the Russian platform, with its crystalline basement complex and undisturbed mantle, germanotype tectonics is manifested on the epiBaikalian craton of Timan, with the gently sloping brachyform folds of the mantle characteristic of it, complicated by low-amplitude (up to 500 m) upthrusts and steep overthrusts. In their general features, mantle structures inherit the Baikalian plan and even preserve its southwestern vergence, though in a weakened form. Relatively young (Early-Paleozoic) transverse breaks are widely manifested, which are connected with the development of negative structures separating

the dome-block uplifts of Northern, Central, and Southern Timan. Manifestations of magmatism of the platform stage are represented by Early-Paleozoic dikes of picrites with carbonates in Central Timan [7], minor stocks of gabbros with copper-nickel mineralization [4], and lamprophyres in Northern Timan, nappes of Late-Devonian trapps and associated pyroclastic rocks traced along the whole of Timan, and also with isolated pipes of Devonian kimberlites revealed in Central Timan.

Ordovician (?) deposits, which correlate with the Nibel suite of the Pechora syncline, were preserved in Central Timan, where they are represented by red sandstones and siltstones, not very thick (up to 115 m), containing large titanium placers. The complex of Silurian and Lower-Devonian deposits, which is distinguished only in Northern Timan, lies transgressively on the Baikalian basement complex and minor intrusions of the orogenic stage breaking through it. It is represented by clay-carbonate rocks, not very thick (up to 100 m). In relation to paleoplacer gold and diamonds, Lower-Paleozoic deposits of the Caledonian cycle of sedimentation are of no interest, since they almost do not contain gritstones or conglomerates; the same is true of fine-clastic, clay and carbonate strata of the upper part of the Devonian, and the Carboniferous and Permian, which occur over a widespread area in submerged structures, have considerable thickness, and conclude the transgressively constructed Hercynian tectonic-sedimentation cycle.

From a prospecting point of view, we are mainly interested in highly mature terrigenous deposits of the Middle and lower part of the Upper Devonian, which begin the Hercynian cycle of sedimentation. They lie unconformably on fragments of the mantle of Lower-Paleozoic age and the Baikalian basement complex, with locally preserved bauxite-bearing and hydromica-kaolinite weathering crusts. On carbonate rocks of the Riphean, Devonian deposits fill ancient karst cavities and have a dislocated position connected with repeated collapses of sinkholes and depressions (as in the Upper Tsil'ma basin in the northern part of Central Timan). The deposits combine second-order transgressive sedimentary cycles with quartz-pebble, slightly cemented conglomerates at the base of each one. In contrast to underlying and overlying strata, which initially had a widespread blanket occurrence, the conglomerate-containing formations under consideration are characterized by sharply variable thickness; they are separated one from the other by erosion surfaces and accumulated in shallow (as much as 500 m) grabenlike depressions, which were probably of a riftogenic nature. This is especially true of the lower (Middle-Devonian) formation (with thickness up to 300 m in Northern and up to 415 m in Central Timan), which quickly wedges out to the sides of the depressions. Two overlying formations with thicknesses up to 30 (lower) and 95 m (upper) partially extend beyond the bounds of the depressions and "seal" them.

We established the main direction of transport of ore-bearing clastic material during accumulation of productive formations according to the orientation of large-scale and medium cross-bedding in sandstones and puddings: from west and northwest to east and southeast [13-16]. The feed areas were ledges of the Baikalian basement complex and an extensive part of the Russian platform adjacent to Timan, which remained elevated in the Middle and beginning of the Late Devonian [2, 15].

In conglomerates and gritstones of Middle- and Upper-Devonian age, in Northern, as well as Central Timan [1, 11, 13, 15, 16, 19, 20, 22], a large number of occurrences of placer gold have been established, which is often accompanied by diamonds. Gold has also been discovered recently in Devonian quartz conglomerates in the continuation of Timan, the Kanin Peninsula (oral communication of A. G. Valasis). In Southern Timan, where corresponding investigations have not been conducted, gold has not been established in Devonian conglomerates, although its presence is highly likely. However, the first diamond crystals were recently discovered in them. Besides gold and diamonds, placer concentrations of niobium minerals have been established in Central Timan: ilmenorutile, columbite, and, in places, pyrochlore, which are accompanied by rare-earth xenotime, monazite, kullerite, and florencite. The quantitative ratios of all of these valuable components vary from place to place, which is due to the location of the corresponding primary sources; occurrences are known with predominance of sometimes gold, sometimes diamonds, and sometimes rare-metal and rare-earth minerals.

The paleoplacers belong to the dynamic classes of nearby (occasionally), moderately nearby, and moderately far-off multistage transfer. Genetically these are, respectively, karst, draw, alluvial—alluvial-fan, delta (small alluvial fans subjected to reworking by the sea), and beach ones.

The gold in conglomerates and gritstones is very fine, mostly 0.1-0.25 mm (Table 1). According to the data of L. A. Nikolaeva and S. V. Yablokova (TsNIGRI), its main peculiarities are deep, repeatedly renewed corrosion, and also signs of repeated redeposition and transportation of gold particles, which are typical of gold from ancient conglomerates in known foreign objects. The gold particles are characterized by zones of translations, recrystallization, and compaction of

TABLE 1. Characteristics of Gold from Primary Sources and Devonian Conglomerates in Northern and Central Timan

Gold-containing rocks	Impurity elements, %							Fineness	Predominant size of gold particles, mm
	As*	Cu	Fe	Mn	Pb*	Pd*	Sb*		
Zone of gold-quartz-sulfide mineralization in carbonaceous shales and meta-aleurolites	0,001-0,009	0,005-0,3	0,02-0,04	0,0001-0,0025	0,0025-0,007	0,003-0,01	0,002-0,007	945-980	0,07-0,5
Conglomerates**	0,001	0,003	0,01	0,0001	0,002	0,002	0,001	955-999	0,1-0,25

Remark. Microspectral analysis of gold was performed at the laboratory of the TsNIGRI (analyst I. P. Lantsev).

*The element is not found in all samples.

** Average content for 12 samples.

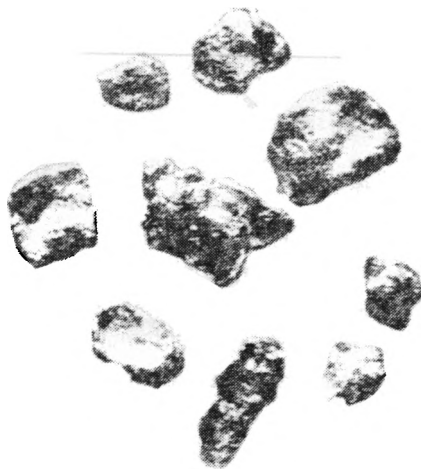
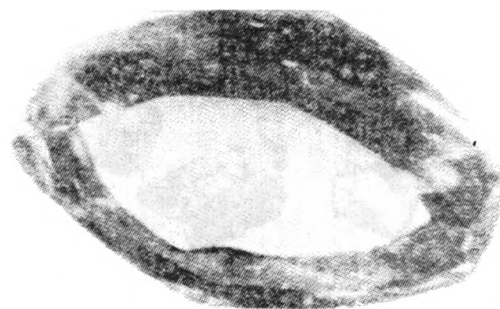


Fig. 2. Placer gold from Devonian quartz conglomerates of Central Timan (mag. 75).

the corrosion layer. The fact that the gold spent time in weathering crusts is confirmed by the presence of several generations of intergrain veinlets. The characteristic strong metallic luster of the surface of the gold particles' corrosion layer is due to its unbroken recrystallization, forming new grains. Less transformed gold particles are found in a subordinate amount, the corrosion of which varies from rudimentary to unbroken occupying from 5 to 30% of their volume. Cases are noted of a build-up of secondary gold on low-grade residual gold. The particles' morphology varies. Toroidal, flattened gold particles with a thickening (ridge) around the edges are found (Fig. 2), and also platy, tabular, lumpy, and dendritic ones. Hemiidiomorphic and distorted crystalline forms are found. The average fineness is high (955-999). Approximately the same values of fineness were established by B. A. Bogatyr et al. (IGEM RAN) in studying gold particles on a "Kameka" MS-40 microprobe. On the surface of the gold particles, kaolinite paste, and films and excrescences of iron and, less often, manganese hydroxides are found, confirming that the gold spent time in chemical weathering crusts. The impurity elements in the gold, revealed by quantitative microspectral analysis at the TsNIGRI laboratory (for 12 samples), are given in Table 1. The association is similar to that in segregations of native gold from hydrothermal zones of vein-disseminated quartz-sulfide and sulfide mineralization localized in carbonaceous strata of the Baikalian basement complex. This allows us to preliminarily consider such zones as the main primary sources of placer gold in Devonian conglomerates.

The diamonds in Devonian placers are comparatively large (average weight of entire crystals 30-50 mg, maximum up to 249 mg), high-quality, primarily water-transparent, and curved-faced. Dodecahedra predominate, which makes them similar to diamonds from Devonian and contemporary placers on the West slope of the Central and Northern Urals and Polyudov Kamen. As on Ural diamonds, green and brown spots of surface pigmentation are found on them. However, in contrast to many Ural diamonds (as from the Krestovozdvizhensk and Vizhai placers), they are practically not rounded, or they only bear traces of alluvial wear: notches on the edges, chips out of the faces (Fig. 3). The action of river flows is also indicated by a significant number of small diamond chips with weight of less than 5-7 mg. All of this allows us to suggest the relative proximity of the primary sources: within the range of 70-100 km. Judging from the presence in the conglomerates of a small amount of accessory mineral (pyrope, chrome-spinelides with Cr_2O_3 content of more than 50% by weight, and distinctive zircon [10]), these sources were kimberlites. However, no diamond-bearing kimberlites have yet been discovered in Timan (the three known pipes on the Vym ridge do not contain diamonds). This is apparently to be expected, since nowhere in the world have productive kimberlites been established in the Baikalian or younger border of ancient cratons. At the same time, diamond-bearing kimberlites have now been revealed on the northern margin of the Russian platform, where they are partially covered by the Carboniferous and probably have a Middle-Paleozoic age. Taking into account the direction of transfer of clastic material, all of this allows us to think that the primary sources of Timan diamonds were diamond-bearing kimberlites mainly located beyond the bounds of Timan, in the adjacent part of the Russian platform. In the contemporary structure, they are buried under a thick platform mantle of the Carboniferous,



a



b

Fig. 3. Diamonds from Devonian quartz conglomerates of Central Timan (mag. 20): a) dodecahedroid without signs of mechanical wear; b) fragment of curved-faced crystal with signs of intensive alluvial wear.

Permian, and younger deposits, which explains the difficulty of discovering them. Still, we cannot rule out the probability of finding diamond-bearing kimberlites in Timan, considering the epochs of riftogenesis that were twice repeated in the Paleozoic, which are favorable for kimberlite volcanism: at the boundary of the Vendian and Cambrian, and in the Middle Paleozoic (end of the Early-beginning of the Middle Devonian). The former is connected with the establishment (or, possibly, rejuvenation) of faults and strike-slip faults transverse to the Timan structure, and the intrusion of clusters of dikes of alkali-ultrabasic rocks with carbonatites, which have been dated as 570-590 million years old. The latter is connected with the formation of numerous, shallow, graben-like depressions and the manifestation first of oreless kimberlite volcanism (the pipes on Vym ridge with a K-Ar age according to phlogopite of 385 million years – N. A. Dovzhikov's data), and then trapp volcanism. Thus, the primary sources of diamonds, from our point of view, are mostly outside of Timan, and only partially in Timan.

In contrast to them, the primary sources of niobium and rare-earth minerals contained in the conglomerates, like the sources of gold, are found in the local Baikalian basement complex. They were the contact parts of dikes of picrites with carbonatites, and also zones of aegirine-feldspar metasomatites, which are known in Chetlasskii Kamen of Central Timan [7]. This determines the comparatively nearby transfer, on the whole, of ilmenorutile, columbite, monazite, kullerite, xenotime, and florencite to the areas of their concentration in conglomerates and gritstones.

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1. The prevalence of "black-shale" formations among Upper-Precambrian strata of the Baikalian basement complex of Timan, the unity of the inner structural zone, which is traced through Southern, Central, and Northern Timan to the

Kanin Peninsula, and the presence in it of numerous gold-ore occurrences in shales, and of different-aged placer occurrences connected with their erosion, enable us to distinguish Timan as a large potential gold-ore zone located in the relative accessible European part of Russia; and ledges of the basement complex of Southern, Central, and Northern Timan, accordingly, as potential gold-ore regions. The latter are promising for discovery of deposits similar to those known in carbonaceous terrigenous-shale formations of the Patom highland, the Sayan region, and the Enisei chain of hills, and also with objects in carbonate strata and zones of oxidation.

2. In potentially ore-bearing regions of Timan, diamond-bearing kimberlites may have been formed. This is indirectly indicated, first of all, by the twice repeated epochs of riftogenesis in the Paleozoic, which are favorable for manifestations of alkali-ultrabasic (including kimberlite) volcanism, and secondly, by the diamonds' slight wear, on the whole, which indicates their nearby transportation (possibly from uplifts of the local basement complex).

3. On the flanks of potential gold-ore regions, where highly mature terrigenous formations of the Devonian are exposed, in conglomerates and clay puddings filling ancient washouts, paleovalleys, and karst depressions, we can expect to find polymineral placers from nearby erosion with variable amounts of gold, diamonds (possibly large ones here), and rare-metal and rare-earth minerals. In association with them, significant concentrations of niobium and rare earths in kaolin clays and bauxites of Devonian age are of prospecting interest [3]. Further from ledges of the Baikalian basement complex to the east and southeast, as the facies change in conglomerate-containing horizons, we can predict stratiform paleoplacers of moderately far-off erosion of the type revealed in the Ichet-Yu area [13, 16, 19, 20, 22].

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EVAPORITES AND THE ORIGIN OF ORE.

COMMUNICATION 2. DISTRIBUTION CHARACTERISTICS AND CRITERIA FOR PREDICTING MINERAGENCY OF POTASSIUM-BEARING FORMATIONS

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Characteristics of the minerageny of potassium-bearing formations and the role of halogenesis (halogenic formations) in development of the set of ore and nonmetallic mineral resources accompanying them are analyzed. Criteria are determined for prognostic evaluation of this whole complex.

MINERALOGICAL CHARACTERISTICS OF POTASSIUM-BEARING FORMATIONS AND THEIR LITHOCOMPLEXES

Among halogenic formations, potassium-bearing ones have the most significant (hundreds and a few thousands of meters) thicknesses and the lithologically most diverse and complete cross sections. They are characterized by rhythmic structure, thanks to alternation of clay, carbonate, anhydrite, and salt rocks. The role of these rocks and the degree of salt saturation of the cross sections of specific formations vary considerably. Potassium-bearing formations more often have the most salt-saturated cross sections and include the most complete sets of salt rocks, including various potassium and magnesium salts, and also microcomponents accompanying them, such as bromine, boron, rubidium, cesium, etc.

With respect to the set of potassium and magnesium salts, potassium-bearing formations, as was already noted above, are divided into three types: chloride, sulfate-chloride, and sulfate. In the first type of formations, three subtypes are distinguished: sylvinite-halite, sylvinite-carnallite-halite, and sylvinite-carnallite-bischofite (tachhydrite)-halite. Each of these types and subtypes of potassium-bearing formations, besides differences in the mineral composition of potassium and magnesium salts, also has other peculiarities, including lithology and stratiform (superimposed) mineralization [1, 2].

The most common in time and space are potassium-bearing formations of the chloride type, in which deposits of potassium and magnesium salts are represented by chlorides only: sylvinite, carnallite, and bischofite (or tachhydrite) rocks. The three subtypes of these formations mentioned above differ in the completeness of the set of these rocks: from sylvinite to sylvinite-carnallite to sylvinite-carnallite-bischofite (or tachhydrite). The most common of these three subtypes of formations is sylvinite-carnallite-halite, with a wide time range from the Cambrian to the Paleogene. The subtype with the complete set of potassium and magnesium salts is characteristic only of Lower-Cretaceous formations in Brasil, west Africa, and Thailand and Laos. There are few formations of the sylvinite-halite subtype; they are not very large in scale and are primarily connected with younger Cenozoic deposits.

Potassium-bearing formations of the sulfate-chloride type, along with the predominant deposits of chloride potassium and magnesium salts (the same as in formations of the chloride type), include sulfate potassium and magnesium rocks in less significant volumes: polyhalite, kainite, langbeinite, kieserite, etc. The lithology of these formations' cross sections and the sets of rocks surrounding deposits of potassium and magnesium salts differ little from such potassium-bearing formations of the chloride type. One peculiarity of them alone is the rocks' boron content, which frequently reaches commercial value (at least for by-product extraction). The cation composition of borates in these formations' underlying rocks is determined by the lithology of boron-bearing horizons. When these horizons reach the zone of hypergenesis, more boron-enriched deposits of residual borates are formed, which are more promising and accessible for exploitation. Potassium-bearing formations of the sulfate-chloride type are characteristic of the Permian.

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In chloride potassium and magnesium salts of sulfate-chloride potassium-bearing formations, as well as chloride ones, the presence of high bromine content has long been established, reaching commercial values in carnallite and, especially, in bischofite rocks. In recent decades, horizons with increased contents of rubidium (lower carnallite layers) and sometimes cesium (concluding carnallite layers) have been discovered. These components can be extracted along with exploitation of surrounding potassium and magnesium salts.

In potassium-bearing formations of the chloride and sulfate-chloride types, the valuable mineral raw materials for the mining industry are primarily rock salt and potassium and magnesium salts, and also components associated with them: borates, bromine, rubidium, and cesium. Salts are consumed in especially large amounts in the economy. Table salt is essential to people. Like potassium and magnesium salts, it is consumed in the economy in enormous amounts.

Besides the traditional commercial salt material, certain other useful minerals are connected with anhydrite-carbonate-terrigenous horizons and lithocomplexes of these two types of potassium-bearing formations. First of all, they are the rocks themselves of these horizons, such as gypsum (anhydrite) and various carbonates (limestones, less often dolomites, and even magnesite). They are also characterized by the presence of deposits of native sulfur. They were formed especially often in regions where hydrocarbons were present in deposits underlying potassium-bearing formations. Sulfur formation was promoted by the proximity of potassium-bearing formations (a barrier) to oil and gas fields, and the presence of disjunctive tectonics, which resulted in reduction of sulfates to hydrogen sulfide by hydrocarbons, and then to native sulfur. The latter was deposited in carbonate horizons contacting anhydrites or gypsums. In such conditions, deposits of native sulfur were formed in carbonate-gypsum caprocks of salt-dome structures.

In peripheral anhydrite-carbonate lithocomplexes of potassium-bearing formations, celestine (sometimes barite-celestine) layers are fairly often represented, and sometimes even fluorite-containing ones. Celestine-containing horizons are present in almost all potassium-bearing formations that contain anhydrite-carbonate rocks. If horizons of anhydrite and carbonate rocks alternate repeatedly, then there may be several such celestine-containing interlayers in the formation's cross section. Fluorite-containing layers are more often only of mineralogical interest, due to their low (15-25%) fluorite content. In individual regions (for example, in Permian formations of the Kungurian and Zechstein), they are richer in fluorite (30-50%).

In places, carbonate rocks of peripheral lithocomplexes of the types of formations indicated above are connected with sulfide lead-zinc mineralization, and sometimes mercury and manganese ore content. In terrigenous rocks directly surrounding (underlying) such potassium-bearing formations, horizons of cupriferous sandstones or shales are often found. In some halogenic regions, they are copper ores with commercial value.

Potassium-bearing formations of the sulfate type have a specific lithology and a younger, Cenozoic age. They are characterized by high clay content of salt rocks and of the formations' whole cross section. Horizons of carbonate rocks are often almost completely absent in their cross section, while anhydrites are represented by isolated, thin layers. Deposits of potassium and magnesium minerals are primarily composed of such sulfate-magnesium minerals as kainite, langbeinite, kieserite, and polyhalite. Of chlorides, sylvinite and, less often, carnallite are represented in them. Thanks to the high clay content of rocks in the cross section of formations of this type, their boron content is extremely low. As a consequence of low carnallite content, the prospects for increased contents of bromine and rubidium in rocks of the cross section are not good. Finally, due to the absence of horizons of carbonate rocks in potassium-bearing formations of this type and their poverty of anhydrite layers in the potassium-bearing subformations proper, lean horizons with inlets of native sulfur and lead-zinc sulfide mineralization are very rarely found. Richer ores of this type are sometimes found on the periphery of these potassium-bearing regions, where they are linked with carbonate-anhydrite (gypsum) strata underlying or overlying these formations and, at times, not directly connected with them.

As we can see, potassium-bearing formations of each of the three types have their own distinct specialization in the set of potassium and magnesium salts and surrounding rocks, and partly also in their accompanying ore and nonore mineralization. Such differentiation of these cross sections' lithology and mineral composition occurs thanks to a regular change in the physicochemical conditions of sedimentation in potassium-bearing halogenic basins of different types. Their classification was determined by evolution of the composition of the World Ocean's water in time, and by the lithology and chemism of waters in the areas of denudation.

The most common potassium-bearing formations of the chloride types are present among deposits of the Cambrian, Devonian, Permian, Triassic, Cretaceous, and Paleogene. Potassium-bearing formations of the sulfate-chloride type are characteristic of the Permian; and formations of the sulfate type, of the Cenozoic. We will also note that the sylvinite-carnallite-bischofite (tachhydrite) subtype of potassium-bearing formations of the chloride type, which is characteristic of

Lower-Cretaceous formations of Brasil, Congo, and Thailand, is rather uncommon and distinctive. The classification of potassium-bearing formations, which is determined by the mineral composition of potassium and magnesium salts, enables us, in turn, to easily predict the prospects for increased contents of a number of accompanying components in them, such as bromine, rubidium, and cesium. The higher magnesium content of rocks (dolomites and even interlayers of magnesite, carnallites, and bischofites) in potassium-bearing formations of the sulfate-chloride type, with low clay content of halogenic rocks in their cross section, created favorable conditions so that it is precisely formations of this type that are connected with specific, fairly significant boron content of salt rocks, as well as of carbonate-gypsum caprocks.

The presence of stratiform and superimposed ore and nonore mineralization discovered in peripheral anhydrite-carbonate-terrigenous lithocomplexes of potassium-bearing formations is determined by their lithology, the intensity of disjunctive tectonics, and the presence of hydrocarbons in subsalt deposits. The presence of ash material in terrigenous interlayers of some formations does not rule out the entry into halogenic basins of ore-bearing hydrothermal solutions similar to those that presently enter depressions in the bottom of the Red Sea.

Already in the first stages of salination of basins (precipitation of sulfates and mass death of organic matter), zones of hydrogen-sulfide contamination arose, which created hydrochemical barriers at which ore components entering the basins were caught and precipitated. It is precisely in such conditions, in the lowest part of the cross sections of halogenic formations, and in their peripheral lithocomplexes or adjacent lateral equivalents, that horizons of cupriferous sandstones or shales could form, and then carbonate rocks with sulfide lead-zinc mineralization. Such were the conditions for concentration of lead-zinc mineralization in subsalt carbonate horizons or carbonate reefs bordering halogenic formations. The favorable situation for ore formation in all of these cases was created by halogenesis: removal of ore material by chloride brines (its solvents) and precipitation of it at sulfate (hydrogen-sulfide) barriers formed by halogens. This indicates the close connection of halogenesis and the origin of ore in potassium-bearing regions.

Subsequently, in the main phase of accumulation of salt strata, the latter usually became reliable barriers for subsalt ore-bearing solutions and hydrocarbons. However, in peripheral belts of salt-bearing basins, where the thickness of halogenic formations is not great, and they are represented primarily by anhydrite-carbonate-terrigenous rocks broken up by faults into blocks, in places ore-bearing solutions could enter into the basin's brine and sediments, fixing lead-zinc, celestine, barite, fluorite, and manganese mineralization in some carbonate horizons and again at sulfate (hydrogen-sulfide) barriers.

In the concluding stage of halogenesis and subsequently, in covering carbonate-sulfate strata or in caprocks of salt-dome structures conditions also arose for formation of lead-zinc mineralization, and sometimes fluorite, barite, and celestine. Processes of diagenesis redistributed the mineralization of stratiform celestine- and fluorite-containing (ratofkite) horizons.

Thus, the sets of mineralization in different lithocomplexes of potassium-bearing formations are less often differentiated than in specific types and subtypes of these formations. In connection with this, there is increased interest in revealing the patterns of distribution and succession of various lithocomplexes of potassium-bearing formations of all types and subtypes, and also the sets of mineralization of each lithocomplex, and revealing the clear stratification and confinement of mineralized horizons to certain varieties of rocks.

Over a significant part of their area, potassium-bearing formations are composed primarily of salt rocks forming a generalized (basic, overall) salt lithocomplex. Its thickness drops sharply to the rims (faults), beyond the bounds of which it extends only locally and comparatively slightly. The salt lithocomplex is bordered by anhydrite-carbonate-terrigenous lithocomplexes. In the cross section of potassium-bearing formations, the role of layers of first anhydrite, then carbonate rocks and, finally, terrigenous ones rises sharply from their center to the periphery, beyond the bounds of the rims. Such is the broadest fundamental arrangement of potassium-bearing formations' lithocomplexes.

The salt lithocomplex proper (halite-containing) is differentiated into anhydrite-halite and primarily halite lithocomplexes. Each type and subtype of potassium-bearing formations is represented by specific sets of potassium-bearing lithocomplexes, which is determined by their characteristic stratification of layers of different potassium and magnesium salts, and by changes in the areas of their occurrence at each level, with the prevailing trend of their reduction toward the upper part of the cross section. Local anomalies in the distribution of potassium-bearing lithocomplexes are due to manifestations of local tectonics (various uplifts, block structures, ledges, etc.).

Distribution characteristics (rather regular) of potassium-bearing lithocomplexes are most clearly revealed visually for potassium-bearing formations of the chloride (and also for their subtypes) and sulfate-chloride types. Their arrangement is the simplest for potassium-bearing formations. In formations of the sylvinite-halite subtype, there is only one potassium-

bearing lithocomplex, with the same name as the subtype. For the carnallite-sylvinite-halite subtype two lithocomplexes are distinguished: the sylvinite-halite lithocomplex is often encircled by a carnallite-sylvinite-halite one. Finally, the bischofite-carnallite-sylvinite-halite subtype is characterized by the presence of three lithocomplexes. Their zonal location is disrupted only in potassium-bearing structures with broken block structure and in connection with intensive formation of zones of substitution and dilution in deposits of potassium and magnesium salts.

Potassium-bearing formations of the sulfate-chloride type are characterized by the presence of the same three potassium-bearing lithocomplexes as in chloride potassium-bearing formations, with the most complete sets of potassium and magnesium salts. These lithocomplexes, which are located in the most depressed central part of a halogenic basin, locally include subordinate interlayers of sulfate potassium and magnesium rocks with kainite, langbeinite, or kieserite. Only in isolated formations do these rocks reach practical significance and form special lithocomplexes. In peripheral bands on the edges of salt lithocomplexes, along with sylvinites, layers of polyhalite rock, sometimes very thick, are represented in places. There are sylvinite-polyhalite-halite, as well as polyhalite-halite lithocomplexes here. The carnallite and, especially, bischofite rocks of these formations are richest in bromine; and carnallite rocks, also in rubidium and, less often, cesium. Potassium-bearing formations of the sulfate-chloride type are characterized by boron content in rocks of their cross section.

Almost unique potassium-bearing formations of the sulfate type have the most variegated patterns of succession of potassium-bearing lithocomplexes. Here, they are usually localized in small depressions of the most mobile transverse blocks of the basement complex. In the axial or central part of the basin, lithocomplexes containing kainite and langbeinite rocks are predominant, and also rocks of a more complex composition: of the hartsalz type. The quantitative composition of these rocks in layers is subject to sharp changes. Hartsalz and langbeinite-kainite-containing, and also mixed lithocomplexes are represented here. But distinguishing them is a rather difficult problem.

In peripheral sections of sulfate potassium-bearing formations near the platform, along with kainite rocks, there are deposits of sylvinites in places and, less often, carnallite rock. In these areas, a kainite-sylvinite lithocomplex is distinguished. In Ciscarpathia, it is represented by clay salt rocks.

Peripheral anhydrite (gypsum)-carbonate-terrigenous lithocomplexes of potassium-bearing formations may be of interest as possible deposits of gypsum and carbonate rocks. The carbonate horizons of anhydrite-carbonate lithocomplexes are connected with large deposits of native sulfur, and with ore occurrences of celestine, fluorite, lead and zinc sulfides, and manganese mineralization. The concentration of this set of mineralization is determined by peculiarities of the cross-sectional lithology of peripheral lithocomplexes and by geological-structural factors (the presence of faults, uplifts), and also by the presence in subsalt deposits of hydrocarbons, which reduce natural sulfates to hydrogen sulfide and create hydrochemical (hydrogen-sulfide) barriers at which the mineralization indicated above is caught. Transfer of ore components is accomplished by chloride solutions, in which their solubility increased. The close interrelation of halogenesis and the origin of ore determines the regular distribution of ore and nonore mineralization in certain lithological horizons and lithocomplexes of halogenic formations. This interrelation existed in the stages of sedimentogenesis, diagenesis, and even hypergenesis.

When a weathering crust of halogenic formations develops (so-called clay carbonate-gypsum caps and caprocks), readily soluble salts are removed from halogenic underlying rocks, primarily by processes of hypergenesis, and brine horizons form. Among insoluble residual materials, the concentrations of ore and nonore mineralization accompanying halogenic formations rise accordingly: borates, celestine, barite, fluorite, and polymetallic sulfides. Gypsified layers of anhydrite become mineral material more favorable for exploitation. Under carbonate-gypsum and clay-gypsum caps over the heads of layers of sulfate potassium and magnesium salts, deposits of mirabilite form; and over them, horizons of mineral waters.

CRITERIA FOR COMPLEX PROGNOSTIC EVALUATION OF HALOGENIC REGIONS FOR MINERAL SALTS AND ACCOMPANYING NONMETALLIC AND ORE MINERAL RESOURCES

Prognostic evaluation for mineral salts is presently carried out with sufficient reliability according to the sum of positive indications of a number of prospecting criteria. Among them, the decisive ones are tectonic (structural) criteria, since halogenic formations are confined to large, intensively depressed elements of the Earth's crust. Such are marginal

troughs at the edges of platforms, marginal troughs of intermontane depressions, and intraplatform aulacogens, and also synclises and grabens superimposed on them. The role of stratigraphic (epochs of intensive salt accumulation), lithological, paleoclimatic, paleogeographic, and hydrogeochemical criteria is significant for prediction [3].

Within the bounds of the troughs, synclises, and aulacogens indicated above, the tectonic criterion distinguishes the most depressed areas of basins and transverse blocks. They are the most promising for the presence of potassium and magnesium salts. Structural maps of halogenic formations (isothicknesses, isohypses of foot and the roof) and the subsalt basement complex enable one to make this evaluation fairly detailed, for the thicknesses of halogenic formations are maximum and the cross sections most complete, as a rule, in intensively depressed areas of these structure, and the prospects for their potassium content are also optimum. The completeness of the formation's cross section is directly related to the degree of diversity of the set of potassium and magnesium salts in it.

Potassium-bearing formations that screen accumulations of hydrocarbons, thanks to the block structure of the subsalt basement complex, are promising for creation of hydrogen-sulfide barriers and deposits of native sulfur and polymetallic sulfides. The most favorable for this are zones near the sides of linear faults, and also sections where they are intersected by transverse faults.

The faults indicated above are also a factor promoting transportation and deposition of fluorite, celestine, and barite mineralization. These mineralized solutions formed vein, as well as stratiform (lenticular) ore accumulations.

Structural tectonic maps of surfaces of the basement complex and the subsalt bed, and also of disjunctive dislocations in subsalt and salt-bearing strata, are used for prediction. In halogenic regions that have been studied in greater detail, maps of the isothicknesses of potassium-bearing formations and a series of geological cross sections are compiled for this purpose, on which dislocations that have been revealed and their accompanying mineralization are noted.

Usually, large (in area and thickness) halogenic formations confined primarily to troughs bordering platforms and to intraplatform grabens are potassium-bearing. These structures are bounded by extended linear boundary faults, near which the interface of salt and saltless lithocomplexes of the formations indicated above usually passes. In potassium-bearing formations, postsedimentation tectonics was manifested in the development of folding, overthrust dislocations, and faults all the way to creation of salt-dome and diapir structures. High tectonic activity in potassium-bearing regions is precisely what caused the creation of favorable conditions for precipitation of a more complete set of mineral salts, and also components accompanying them, as well as ore and nonore mineralization connected with anhydrite-carbonate horizons of the given formations and with caprocks of salt domes.

The stratigraphic criterion, or consideration of epochs of intensive salt accumulation connected with development of large potassium-bearing formations, orients geologists to the expediency of priority study of the minerageny of Early-Cambrian, Middle- and Upper-Devonian, Permian, Triassic, Late-Jurassic, Early-Cretaceous, and Paleogene-Neogene potassium-bearing formations. This criterion distinguishes formations that deserve the most attentive study according to their minerageny.

The lithological criterion, following the tectonic one, is one of the most important for mineragenic research in potassium-bearing regions. With its help, regions promising for occurrence of a set of mineral salts and components accompanying them are delineated from areas of occurrence of anhydrite-carbonate strata and horizons with ore and nonore mineralization. In the first approximation, this is achieved by compiling maps of the extent of lithocomplexes, and then of their fractional varieties. The areas distinguished according to these maps as promising for various mineral resources and components are more precisely defined according to data from study of the cross section of potassium-bearing formations, and their location is detailed vertically and littorally.

The typification of potassium-bearing formations that was carried out and compilation of maps of the extent of lithocomplexes for each type and subtype of them made it possible to see the nature of distribution of various potassium and magnesium salts. This is also tied in with determination of the prospects for distribution of components accompanying them: borates, bromine, rubidium, etc.

Analysis of the distribution of ore mineralization in rocks of the saltless lithocomplex revealed the nature of zonality of the distribution of copper (in terrigenous deposits), zinc, and lead compounds (in carbonate rocks), respectively, from the bottom up vertically and along the littoral from the periphery to the center of the formation. Horizons of cupriferous sandstones and shales are found not only at the foot of halogenic formations, but also in strata immediately surrounding them, often even in areas adjacent to halogenic regions. In places, vein lead and zinc ores with inverse vertical zonality are also discovered in subsalt carbonate strata. At times, barite-cinnabar mineralization is noted in paragenesis with them.

Celestine (sometimes with barite), fluorite, and manganese mineralizations are connected with carbonate horizons of potassium-bearing formations, usually within the bounds of occurrence of their saltless lithocomplexes, often in zones adjacent to edge faults (steps) of depressed structures. It is precisely in these zones that conditions were the most favorable for entry of the corresponding mineralization into a halogenic basin, as well as, later, into its bottom deposits from plutonic sources.

Celestine-barite, fluorite, and also lead–zinc mineralizations are discovered at times in caprocks of salt-dome structures, or rather in their carbonate rocks. Carbonate-gypsum caprocks are therefore promising for prospecting for the indicated mineralization, and also for deposits of native sulfur. These prospects are higher in the edge zones of caprocks, where conditions were favorable for the entry of mineralized waters with hydrocarbons, and for formation of sulfate–hydrogen-sulfide barriers (reduction of sulfates).

The paleoclimatic criterion adjusts the choice of tectonic structures (or parts of them) favorable for intensive halogenesis, since salt accumulation requires an arid or at least semiarid climate. A combination of positive prospects determined by the tectonic, paleoclimatic, and lithological (or at least stratigraphic) criteria makes it possible to resolve the basic problems of predicting halogenic formations and potassium and magnesium salts. The whole set of problems of minerageny of potassium-bearing formations is analyzed with specific consideration of their typification, the presence of carbonate-anhydrite horizons in the cross section, the intensity of disjunctive tectonics, and the presence of hydrocarbons in subsalt deposits.

The paleogeographic criterion, which uses data from reconstruction of potassium-bearing paleobasins (lithofacies, the region of supply with seawater, and runoff from adjacent dry land, etc.), enables one to clarify the basic features of sedimentation and the basin's evolution during salt accumulation, changes in its aquatory (underwater territory), the completeness of cross sections, and their potassium saturation in different parts of the basin. This makes it possible to introduce adjustments to the degree of potassium saturation in steep parts of the basin, where the formations' thicknesses are not maximum. This criterion reveals or scientifically substantiates anomalies in the distribution, or a shift of lithocomplexes or their isothicknesses. Paleogeographic criteria are subsidiary and refining for prediction of potassium-bearing formations.

In potassium bearing regions that have not been adequately studied and explored, additional information about potassium and magnesium salts, the presence of their sulfate varieties, and about boron content and bromine is provided by hydrochemical investigations. A set of special coefficients obtained by scaling of data from chemical analyses of brines in natural springs or wells in the territory of halogenic regions is used as hydrochemical criteria. Such coefficients are: potassic, magnesian, bromine-chlorine, and boric. According to their values, they determine as a result of dissolution of what salts the brines were formed. The hydrochemical criterion offers additional data for prediction of potassium and magnesium salts (different mineral forms of them), borates, and bromine. According to the density of the brines and the height of the local head of groundwater, one can roughly calculate from what depth the brine is carried up, i.e., determine the depth of occurrence of the roof of salts. Salt springs often emerge along faults and, consequently, in turn, trace them. Consideration of hydrochemical data is important not only for prediction of potassium and magnesium salts and borates, but also for evaluation of the water encroachment and dislocation of strata above the salts, which is important for the safety of subsequent exploitation.

The most thorough data for evaluating the prospects for potassium content and the minerageny of halogenic regions are obtained by conducting special geological and geochemical investigations, and prospecting or exploratory drilling for a set of mineral salts and accompanying mineral resources and components.

MOST IMPORTANT FEATURES OF THE MINERAGENY OF POTASSIUM-BEARING FORMATIONS

Increased attention to mineral resources connected in one way or another with halogenic formations has enabled us to see that they are not only a source of traditional salt material, but also contain a number of ore and nonore mineral resources. Significant expansion of the set of mineral resources discovered in halogenic formations and the strata directly surrounding them, naturally, increases their practical significance. At the same time, this indicates the significant role of halogenesis in processes of ore formation, which forces us to pay attention to solution of scientific aspects of this problem. Thus, on the basis of these developments, halogenic regions (and potassium-bearing ones in particular) should be con-

sidered as objects of comprehensive study and prospecting for diverse mineral resources (including ones previously considered nontraditional for them). Along with salt material, this set includes native sulfur, borates, bromine, rubidium, cesium, lithium, celestine (with barite), fluorite, lead-zinc and copper ores, manganese compounds, etc.

Among halogenic formations, potassium-bearing ones have the most diverse and abundant minerageny. The typification and study of these formations' minerageny that have been conducted now make it possible to more reliably predict the mineral composition of specific formations' salts according to a number of data. The classification of a potassium-bearing formation can be foretold at times even when information about its potassium content is still very scanty. The classification of a formation is predicted according to the nature of the lithology of underlying deposits: carbonate rocks underlie chloride potassium-bearing formations; and terrigenous rocks underlie sulfate ones. To predict the type of formation, its age can be used, since the sulfate-chloride type of formation is characteristic only of the Permian; the sulfate type, of the Neogene; and the bischofite-carnallite-sylvinitic subtype of the chloride type, for the Early Cretaceous. In addition, hydrochemical data can be employed to clarify the classification of potassium-bearing formations.

The possibility of predicting the classification of potassium-bearing formations provides the opportunity to evaluate the set of potassium and magnesium salts that will be found in it. The series of lithocomplexes revealed for different types of formations and the characteristics of their distribution, which are controlled by changes in the formation's isothermities and the nature of its overall structure, make it possible to roughly evaluate the areal distribution of various potassium and magnesium salts. In addition, postsedimentation changes in the mineral composition of potassium deposits are taken into account in doing so: the possibilities of substitution and dilution processes, and formation of a "leaching edge."

Prognostic evaluation of boron content, and the presence of bromine, rubidium, and cesium, which are closely connected with mineral salts, is done with consideration of their confinement to certain potassic and magnesian rocks or types of potassium-bearing formations. Commercial boron content is usually connected with potassium-bearing formations of the sulfate-chloride type and with caprocks of their salt domes. Carnallite and bischofite rocks are enriched with bromine; and only certain carnallite rocks, with rubidium and cesium (in lower and upper carnallite layers, respectively).

The rest of the mineral resources of potassium-bearing formations are connected not with salts, but with their saltless lithocomplexes. Usually, these are peripheral anhydrite-carbonate lithocomplexes including deposits of native sulfur, and horizons with celestine (at times, with barite) and fluorite. Less often, intersalt horizons also include fluorite or celestine mineralization.

Of this set of mineral resources, deposits of native sulfur are more significant in scale; they are found in many large potassium-bearing formations of the chloride and sulfate-chloride types, especially in regions where they screen or screened oil and gas fields. Sulfuring of carbonate rocks is confined to fault zones. When horizons of anhydrite (gypsum) and carbonate rocks alternate in these lithological complexes, sulfuring of the latter can be layered. Occurrences and deposits of native sulfur are also present in gypsum-carbonate caprocks of salt domes.

Celestine-containing horizons are widespread in sub-, inter-, and suprasalt carbonate rocks near their contacts with anhydrites. They are known at the foot and roof of sulfured carbonate members. They are also found in anhydrite-carbonate caprocks of salt domes. In formations containing polymetallic mineralization, they are more often composed of barite-celestine and even barite.

Fluorite-containing horizons are found in carbonate members of intersalt and peripheral anhydrite-carbonate strata only in a few potassium-bearing formations. Fluoritization of rocks is confined to fault zones and is locally manifested. In subsalt carbonate strata, fluorite and barite sometimes form veins along faults.

Sulfide lead-zinc and, less often, barite-mercury mineralization is confined to carbonate (sometimes gypsum-containing) rocks of sub-, inter-, and suprasalt potassium-bearing formations. This mineralization is also found in gypsum-anhydrite-carbonate caprocks of salt domes. Upward through the cross section from the foot of the formation and laterally from its boundaries toward the center, zinc mineralization is replaced by lead. Vein ores of subsalt carbonate strata are characterized by inverse zonality of the distribution of lead and zinc ores.

Cupriferous carbonate-containing shales and sandstones are found at the foot of potassium-bearing formations, in the strata surrounding them, and also in deposits replacing them along the littoral and then changing to carbonaceous facies. The copper-mining industry is now based in peripheral bands of such formations and their borders, for example, on the southern border of the Zechstein formation in Central and Eastern Europe.

In peripheral anhydrite-carbonate lithocomplexes of some potassium-bearing formations there is lean manganese mineralization, which is confined to calcareous rocks (gypsums, marls, limestones). The scale of these ore occurrences is not great.

Expansion of ideas about the set of mineral resources connected with halogenic and especially with potassium-bearing formations undoubtedly increases their practical significance. This circumstance, in parallel, led to a recognition of the definite action of halogenesis on formation of the indicated set of mineral resources. Halogens actively promoted the extraction, transfer, and deposition of ore material.

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1. Halogens play an important role in processes of ore formation, hydrothermal, as well as stratiform. Chlorides actively participate in extraction and migration of ore materials from deep underground, and sulfides create hydrochemical barriers at which they are caught, forming vein, as well as stratiform ore deposits.

2. In halogenic regions, ore content is most often confined to their peripheral areas (in subsalt deposits) or to bands bordering them. In halogenic strata, ore content is connected with sub-, inter-, and suprasalt carbonate and terrigenous horizons.

3. The most significant ore content is revealed in potassium-bearing regions, since thicker potassium-bearing formations are confined to intensively depressed structures of the Earth's crust bordered by linear faults. The faults were good paths for the ascent of ore-bearing, chlorine-containing brines, and thick potassium-bearing formations were reliable barriers for them and for precipitation of ore material at sulfate (hydrogen-sulfide) hydrochemical barriers in anhydrite-carbonate lithocomplexes.

4. Development of thick potassium-bearing formations in intensively depressed structures within the bounds of which the subsalt and crystalline basement complex was significantly broken up by faults into blocks was accompanied by the involvement of a significant mass of halogens in processes of ore formation.

5. Larger ore deposits connected with potassium-bearing regions are confined therefore to deposits from periods of the most intensive halogenesis: Early-Cambrian, Middle- and Late-Devonian, Permian, Triassic, Late-Jurassic, Early Cretaceous, and Paleogene-Neogene. Even in intensively metamorphized ore-bearing Precambrian strata, traces of former halogenesis were preserved.

6. The ore-forming role of potassium-bearing formations rises even more in connection with the fact that they are good barriers for hydrocarbons, under the action of which on sulfates hydrogen sulfide and even native sulfur are formed.

7. For halogenic regions, the most common ores are lead-zinc and copper. Besides the set of salts, the most important nonore resources are deposits of native sulfur, borates, fluorite, celestine, and, less often, barite and rhodsite-asbestos, manganese ores, and also such microcomponents as bromine, rubidium, and cesium.

8. Salt strata are connected with borates and microcomponents: bromine, rubidium, and cesium. Native sulfur, celestine, and fluorite are confined to their anhydrite-carbonate horizons. Carbonate horizons of sub- and intersalt strata are connected with lead-zinc and manganese mineralization. Cupriferous sandstones and shales are found in strata of halogenic formations or in lower-lying strata.

9. Potassium-bearing regions in which formations of the chloride and sulfate-chloride types are represented have more significant sets of ore and nonore mineral resources accompanying them than those in which potassium-bearing formations of the sulfate type are developed.

10. The most significant deposits of copper ores and borates are connected with Permian potassium-bearing formations of the sulfate-chloride type and with approximately equal-aged strata in areas adjacent to them.

11. Concentrations of such microcomponents as rubidium, cesium and, less often, lithium sufficient for by-product extraction are revealed in potassium-bearing formations of the chloride and sulfate-chloride types, with the first two microcomponents being connected with carnallite rocks. The highest concentrations of bromine are found in bischofite and carnallite rocks.

12. In salt-dome regions where the salt strata are intensively crushed into disharmonic folds, deposits of potassium and magnesium salts, boron-bearing ones among them, and also including bromine, rubidium, and cesium, are complexly dislocated, in places even boudinaged. In these structures, ore mineralization is found in carbonate rocks near the edges and in suprasalt carbonate-gypsum caprocks. In the latter, deposits of native sulfur are also found, and sometimes barite and residual borates.

13. The ore-forming role of halogenic formations is frequently manifested at some distance from them, beyond the bounds of halogenic regions, since chloride brines in zones of tension could sink all the way to the crystalline basement and be heated, and also migrate laterally through water-permeable horizons.

14. The narrow specialization that has been practiced in our country of conducting geological-survey works in potassium-bearing regions for just one type of mineral material out of the diverse set of mineral resources present in potassium-bearing formations and strata immediately surrounding them has had a negative effect on the completeness and speed of revealing the whole set of them, and also on establishing reliable patterns of their spatial distribution and criteria for prospecting work.

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WAYS OF DISTINGUISHING EPIGENETIC TYPES OF GRAY-COLORED ROCKS IN SHEET-INFILTRATION DEPOSITS

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At an infiltration deposit of uranium in Upper-Cretaceous sandy-clay alluvial deposits (Central Kyzylkum), six stages of postsedimentation epigenetic transformations were distinguished. The genesis of altered rocks was determined mostly by mineralogical methods. A new methodological technique is suggested, which calls for analysis of the textural, structural, and mineral correspondence of ferruginous components of gray-colored and oxidized rocks in the region of their contact and makes it possible to typify epigenetic changes in conditions of repeated change in the direction of geochemical processes.

Exogenous epigenetic accumulations of uranium connected with the development of zones of sheet oxidation in unlithified deposits of the sedimentary cover form one of the leading commercial types of deposits (infiltration, "sand-stone") of this mineral resource in many countries of the world. Interest in deposits of infiltration genesis, even with a reduction in demand for uranium in the conditions of conversion, is connected with the possibility of working them by the highly economical method of underground leaching, and also the presence in the ores of a set of useful accessory components: Mo, Se, Re, Sc, and TR.

Since the time of discovery of exogenous uranium deposits in the late 1950's, the question of determining the genesis of epigenetic transformations of the rocks and combining them into epigenetic zonalities has been given a great deal of attention. A large number of investigations by domestic and foreign authors has been devoted to this problem [2, 3, 7, 8, 11, 12, 17-19]. They developed morphological, lithological-facial, mineralogical, hydrogeological, and other methods of resolving this problem.

A significant part of the deposits of this type was formed by a single-stage, unidirectional, sheet-oxidation process connected with the penetration of oxygen-containing artesian waters into permeable gray-colored sandy deposits (for example, objects of the Syrdar'ya, Chu-Sarysui, and Ili provinces [2, 5, 9, 14]). At other deposits, the formation of ore material where zones of sheet limonitization wedge out was accompanied by pre- and postore manifestations of reduction epigenesis (a number of deposits of the Central Kyzylkum province [1, 8, 13, 14]).

The basic and most common way of revealing multistage epigenetic changes is to establish different-aged mineral associations and then map them, reveal such macroscopic traits of geochemical processes with different directions as "inverse color interrelations" in rocks with different permeability, and map their color [7, 8, 11, 13]. Further from the center of reduction, as the number of epigenetic neogeneses (stages) decreases, the mass and species set of accessory minerals diminish. This, and also the primarily loose nature of sandy deposits, significantly limits the possibility of identifying epigenetically transformed rocks and conducting stage-by-stage analysis of mineral formation according to macro- and microscopic traits. In a number of cases, processes of reduction epigenesis destroy the rocks' color traits corresponding to their primary state connected with sedimentation characteristics, as well as with subsequent epigenetic changes, which leads to formation of strata undifferentiated in color. Thus, for example, a light-gray color of sands and sandstones with a whitish tint is characteristic of originally gray-colored rocks with low content of the basic pigmenting component (carbonaceous plant detritus) and of epigenetically oxidized rocks that have undergone gley or sulfide reduction. Considering the convergence of the macroscopic traits of altered rocks, new, additional mineralogical criteria were introduced to delineate them.

As is known, the main distinctive feature of the zone of sheet oxidation is the pseudomorphic nature of oxidation of minerals containing bivalent iron. Neogenic Fe hydroxides completely inherit their textural and structural peculiarities: the distribution in rocks, morphology, dimensions, and the internal structure of aggregates. It is also known that the

degree of oxidation of the rocks' components increases from the frontal part of the profile of sheet oxidation zonality, which corresponds to the subzone of partial and incomplete oxidation, to its back parts – the subzone of complete oxidation [2, 4]. The aggregate state of pseudomorphs of iron hydroxides, primarily on disulfides, changes analogously: from distinct in the frontal parts to completely destroyed, transformed into unformed ocherous aggregates in back parts [4]. However, in a number of deposits we observed a fundamental disparity between iron-containing components of gray-colored and limonitized deposits near their contact with respect to genetic features and the aggregate state of pseudomorphs, and with respect to the degree of oxidation of the rocks' components. Thus, for example, completely oxidized clay balls and interlayers were discovered right where the zone of sheet oxidation wedges out; heavily destroyed pseudomorphs on iron disulfides were found there; and well preserved ones, in its back parts. Cases were also established when limonitized sands containing pseudomorphs of iron hydroxides on pyrite contact gray-colored deposits in which only marcasite is present. Contradicting the patterns of development of sheet oxidation, such cases characterize a geochemically and hydrodynamically opposite process: epigenetic sulfide reduction of oxidized rocks. Thus, analysis of the textural, structural, and mineral correspondence of iron-containing components of gray-colored and oxidized rocks in the region of their contact, which makes it possible to establish the genetic nature of authigenic minerals, was chosen as the basic methodological technique for typifying epigenetic changes.

The most informative diagnostic traits for gray-colored rocks proved to be the morphology, dimensions, and aggregate state of pyrite and marcasite; and for sheet-oxidized ones, the same parameters of pseudomorphs of iron hydroxides on these minerals. As additional traits, we analyzed the condition of minerals of the heavy fraction (ilmenite), clastic grains of biotite and muscovite, carbonate minerals, and carbonaceous plant detritus.

In genetic analysis of postsedimentation transformations, primary attention is given to binocular study of the rocks themselves, and also the light and heavy fractions separated by washing in water. Such fractionation makes it possible to concentrate and thereby significantly facilitate microscopic diagnosis of minerals, determination of their morphology, dimensions, aggregate state, and other traits, and to evaluate the quantitative ratio of minerals. The method developed by the authors for investigating authigenic minerals according to inclusions of them in clastic micas [15] was used to the full extent. The use of transparent and polished sections was limited to study of dense sandstones with carbonate cement and monomineral fractions.

The methodological techniques carried out in the article for distinguishing epigenetically altered rocks are mostly mineralogical. The results obtained with their help completely agree with conclusions obtained by other methods, and the mineralogical traits for each type of change were so consistent that it was subsequently possible, in most cases, to conduct genetic analysis of the rocks according to individual samples, avoiding the traditional mapping method.

The deposit studied by the authors is located in the Central Kyzylkum province. Ore-containing deposits are represented by a set of alluvial-slopewash rocks of the lower rhythm of the Coniacian-Santonian. They lie on partially eroded deposits of the Upper Turonian and are overlaid by alluvial-slopewash formations of the upper rhythm of the Coniacian-Santonian, marine sandy rocks of the Maastrichtian, and thick strata of Paleogene deposits. The ore-containing horizon is composed of sands and sandstones of channel genesis with a gray color at the base, and light-gray at the roof.

Over a significant part of the deposit, rocks of the upper part of the ore-bearing horizons are colored yellowish-brown and rosy-yellow, including unsorted, variable-thickness, sandy-clay rocks of watershed spaces and floodplain layered siltstones. Through the most permeable parts of the cross section, a zone of sheet oxidation is developed, coloring the sandy rocks yellow, red, and rosy-red, with various shades. Wedging-out of the zone is accompanied by uranium mineralization; however, in certain sections the contacts of gray-colored and oxidized rocks are oreless.

In the territory adjacent to the deposit, occurrences of reduction epigenesis near faults are known, which are represented by kaolinitization, silicification, carbonatization, sulfidization, and bituminization of rocks [13]. However, the aureoles of these changes are fairly local: at the studied deposit they are only expressed by formation of iron disulfides.

In the works of a number of researchers [2, 13] who have used traditional methods of genetic interpretation of the colors of surrounding deposits, a gray color of deposits was interpreted unequivocally as primary. Using the methodological techniques given above to study core material from boreholes, the authors of the present article drew a conclusion about the secondary nature of the gray color of a significant part of the rocks lying mostly in the middle and upper parts of the cross section, which is connected with sulfide reduction of preliminarily epigenetically oxidized deposits. We also established the pulsation nature of the inflow of reducing sulfide solutions that interrupted the progressive development of the zone of sheet oxidation and obliterated its frontal part. These processes led to formation of several epigenetic types of

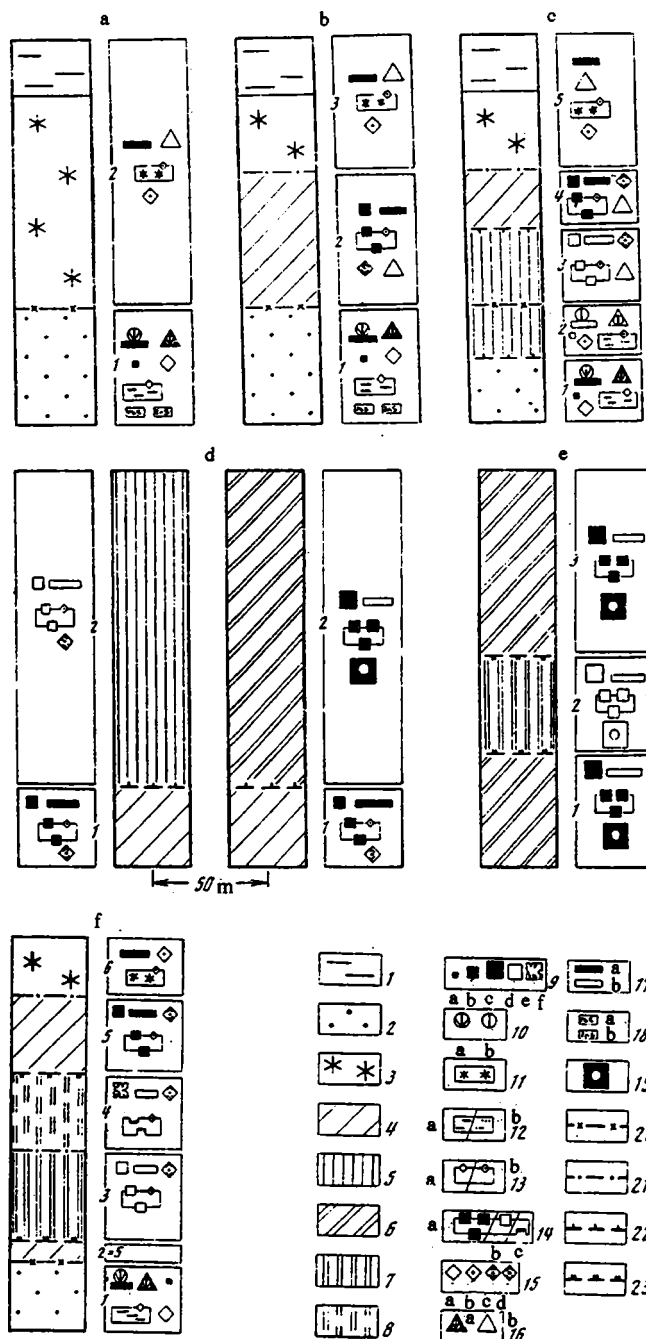


Fig. 1. Variants of interrelations of gray-colored and oxidized rocks in the cross section of ore-bearing deposits, and their mineralogical characteristics (a – variant 1, b – variant 3, c – variant 4, d – variant 6, e – variant 7, f – variant 8). Zones of epigenetically altered rocks: 1) originally red-colored; 2) originally gray-colored; 3) surface-oxidized; 4) secondarily reduced; 5) sheet-oxidized (early zone); 6) repyritized; 7) sheet-oxidized (late zone); 8) postore-reduced; 9-19) authigenic mineralization: 9) crystalline pyrite (a – 5-10 μm , b – 30-70 μm , c – 50-150 μm , d – pseudomorphs of iron hydroxides on ideally preserved pyrite, e – corroded pseudomorphs), 10) framboidal aggregates of pyrite (a – unaltered, b – oxidized), 11-14) mineral inclusions in clastic biotite (a – unaltered, b – oxidized): 11) crystalline goethite, 12) flattened pyrite crystals, 13) rhombohedral crystals of Fe-dolomite, 14) octahedral pyrite crystals (c – relicts of Fe pseudomorphs on pyrite); 15) rhombohedral crystals of Fe-dolomite (a – unaltered, b – with Fe hydroxides in central parts of the rhombohedra, c – with Fe hydroxides and platy pyrite, d – with Fe hydroxides and pseudomorphs of hydroxides on platy pyrite); 16) carbonaceous plant detritus (a – unaltered with framboidal pyrite, b – oxidized); 17) clastic minerals of the heavy fraction (a – unaltered, b – leucoxenized); 18) Pb and Zn sulfides (a – galenite, b – sphalerite); 19) relict forms of oxidation in gray-colored rocks; 20-23) boundaries of epigenetic zones: 20) surface oxidation, 21) reduction, 22-23) early and late sheet oxidation, respectively.

gray-colored rocks and, as a consequence, geochemical barriers differing in their contrast, due to the uneven distribution of mineralization over the area of the deposit. The most typical cases of the interrelation of gray-colored and oxidized rocks are given below, on the example of which a stage-by-stage analysis was conducted.

Systemization of such variants became possible after performing the traditional set of investigations: stratigraphic subdivision of the cross section, lithological and facial analysis of ore-bearing strata with discrimination of lithogenetic types and mineralogical and geochemical characterization of them, determination of the morphology of the zone of oxidized rocks (level or sheet) and the nature of wedging-out of sheet-oxidized zones (wedge-shaped, ledge, ore, oreless), and assembly of a reference collection of unaltered and altered rocks according to their coloring and the nature of oxidation (solid or mottled).

Variant 1. Oxidized rocks of the upper part of the cross section of the ore-bearing horizon (layer 2), including clay water-barrier deposits, contact underlying gray-colored ones (layer 1), which may be represented by deposits of any facial zone of the alluvial complex (Fig. 1a).

In gray and light-gray sands of layer 1, *iron disulfides* are mostly represented by aggregates of framboidal pyrite (2-10 μm , occasionally up to 20-30 μm). The aggregates are localized near carbonaceous detritus, often forming phytomorphs on it, and grow on clastic grains of minerals of the heavy fraction (ilmenite, titanomagnetite, etc.). Sometimes, a crustification radial fringe (2-3 μm) of pyrite and marcasite is observed on the spherical surface of framboids. Solitary octahedral crystals of pyrite (1-5 μm) and short-prismatic ones of marcasite are present in the rocks. The average content of iron disulfides is not great (0.15-0.27%). Near large (> 1-2 cm) fragments of carbonified wood, the amount of iron disulfides and the dimensions of their crystals rise by several times.

A distinctive feature of gray-colored rocks of layer 1 is frequent finds of *galenite* and *sphalerite* in areas of accumulation of carbonaceous detritus. Sphalerite forms fine (5-30 μm), idiomorphic, octahedral and flattened, single crystals and granular aggregates in close association with iron disulfides. The transparent, light-brown to colorless mineral is a low-iron variety — cleiophane. Cubic and octahedral crystals of galenite reach 1-5 mm, are associated with sphalerite, and often form phytomorphs on plant residues.

Fe-dolomite forms individual grains; contact, pore, or basal cement of the crustification type in sandstones; basal coarse-clastic conglomerates; and also poikilitic inclusions in argillaceous cement, interlayers, and balls of silts and clays. In all cases, Fe-dolomite is represented by idiomorphic, rhombohedral, colorless, transparent crystals 30-70 μm in size. According to data from X-ray spectral microprobe analysis (15 determinations), the content of the isomorphic admixture of bivalent Fe averages 2-5%. In the middle part of the crystals, a concentrically closed zone with a higher refractive index is noted, which is connected with an increase (up to 10%) in the content of idiomorphic iron in it. Crystals of Fe-dolomite do not contain pyrite inclusions; the latter is found in the intergranular space of sandstones with carbonate cement in approximately the same amount and with the same morphology as in loose sands. In fragments of pelitomorphic and fine-grained dolomitic rocks from basal horizons, insets of fine-crystalline pyrite (5-10 μm) are noted, which are confined exclusively to the boundaries of carbonate grains.

Organic matter is represented by carbonaceous detritus, partially replaced by framboidal pyrite. According to results of thermal analysis, the organic matter is classified in thermal group 2A, which is characteristic of epigenetically unaltered brown coals [6].

Clastic *hydrobiotite* has a green or deep-green color; its flakes are strong and elastic. Microcavities of cleavage cracks along (001) contain fine (5-10 μm) platy inclusions of pyrite, which, due to their small thickness (0.5-2 μm) do not deform the mica. Clastic *muscovite* is partially hydrated and can contain flattened, fine inclusions of pyrite. Grains of clastic *ilmenite* are black, without traces of change, and aggregates of framboidal pyrite often grow on their surface. The greater part (60-70%) of grains of *clastic potash feldspars* has a gray, dark-gray, or black color, due to inclusions of finely dispersed iron disulfides. Interlayers and balls of silt and clay rocks are gray or dark-gray and contain dispersed carbonaceous detritus and a slight amount of fine-crystalline and framboidal pyrite.

Upward through the cross section, primary gray-colored deposits without any reaction fringes change to yellow, yellowish-brown, and rosy, uniformly colored sands and sandstones of layer 2. Lying at the very roof, slopewash clay deposits have a red or variegated color; layered floodplain siltstones are banded: more clayey varieties are rosy-yellow, and more sandy ones are yellow.

Iron hydroxides in sandy varieties of the rocks are represented by ochers impregnating the argillaceous cement, sandy grains of rock fragments, and feldspars, and cover the surface of all grains of the light fraction in the form of films and clumps. *Dolomite* has the same nature of distribution as in primary gray-colored deposits. Sandstones cemented with

oxidized dolomite are colored rosy or cherry-red, which sharply distinguishes them against the general yellow background of loose rocks. The pigmentation of dolomite is connected with the appearance within it of crystals of zonally located iron hydroxides with a rosy or orange color. The amount and aggregate state of hydroxides vary: dense concentric zones that polish well and scattered, finely dispersed segregations impregnating carbonate material are found. Iron hydroxides originate as a result of oxidation of an isomorphic admixture of bivalent iron present in the dolomite (mostly clearly in the zone with increased content of it), which indicates the epigenetic nature of the coloring of carbonate.

Hydrobiotite is heavily altered; flakes of it are soft and plastic. The mineral's orange-yellow, light-brown, or bronze-yellow color is due to the presence of finely dispersed (1-5 μm) inclusions of crystalline goethite in cleavage microcracks through the flakes' whole volume. The goethite forms feltlike aggregates or numerous, disk-shaped, starlike concretions 5-20 μm in diameter. Analogous, but significantly larger (from 20 up to 250 μm) starlike aggregates of goethite are contained in flakes of clastic *muscovite*. These unusual neogeneses were first established by the authors in rocks in the zone of surface oxidation at one of the uranium deposits [10]. Further study of inclusions of crystalline goethite in clastic micas at more than 10 deposits in other uranium-ore provinces showed that they are a typomorphic trait of rocks of the zone of surface oxidation and originally red-colored rocks [15, 16].

Clastic grains of *feldspars* are impregnated with finely dispersed iron hydroxides and colored yellow, orange, or rosy. Clastic *ilmenite* is visually not leucoxenized. On some grains, round, ocherous clumps remotely reminiscent of pseudomorphs of iron hydroxides on disulfides are found. Similar formations are observed around balls and interlayers of oxidized silty-clay rocks.

Organic matter is almost completely replaced by iron hydroxides devoid of any structural characteristics that would make it possible to establish the morphological type of disulfides that was oxidized. According to results of thermal analysis, the organic matter is classified in thermal group 2B, which is characteristic of oxidized coals [6].

The main macroscopic characteristic of sandy rocks of layer 1 is the facial dependence of their gray color, which, in conjunction with mineralogical traits, the most important of which are the presence of framboidal pyrite, phytomorphs of framboidal pyrite on carbonaceous detritus, sphalerite and galenite near carbonaceous detritus, dark-green clastic hydrobiotite with fine inclusions of crystalline pyrite, and grains of black and dark-gray clastic potash feldspar with finely dispersed iron disulfides, enable us to diagnose the gray-colored rocks of layer 1 as epigenetically unaltered, originally gray-colored.

The morphology of the zone of oxidation in the overlying deposits of layer 2 has a level nature and a uniform type of coloring; there are no reaction fringes or uranium mineralization at the contact with layer 1. Inclusions of crystalline goethite in clastic micas, and structureless, ocherous iron hydroxides in the intergranular space are characteristic of rocks of the zone of the ancient oxidation surface and primary red beds, however the presence of pseudomorphic-oxidized Fe-dolomite rules out an originally red-colored genesis of the deposits. Thus, the rocks of layer 2 have to be considered epigenetically oxidized and classified in the zone of subintermittant surface oxidation. When overlying deposits of the upper rhythm are composed of originally gray-colored rocks, the age of this zone is dated as pre-Upper-Coniacian-Santonian. In some sections of the deposit, where the upper rhythm is also surface-oxidized, penetration of the zone of oxidation from other, higher stratigraphic levels and, accordingly, of a different age is possible.

Variant 2. In the cross section of the lower rhythm of the Coniacian-Santonian aquiferous horizon, between originally gray-colored (layer 1) and surface-oxidized (layer 3) deposits lie light-gray and whitish-gray sands (layer 2, see Fig. 1b). Sometimes, the whole sandy part of the cross section has this coloring, as do clay water-barrier deposits adjacent to it. The contacts of all layers do not contain uranium accumulations.

In the intergranular space of gray-colored rocks of layer 2, *iron disulfides* are primarily represented by idiomorphic, solitary, octahedral crystals of pyrite (30-70, sometimes up to 100 μm) with small and narrow faces of a cube and pentagondodecahedron. Significantly less often, fine (10-50 μm), short-prismatic crystals of *marcasite* are found, sometimes epitaxially growing together with pyrite. The ratio of pyrite to marcasite is equal to 5:1-10:1. The amount of iron disulfides increases in comparison with originally gray-colored rocks to 0.24-0.36%.

Flakes of clastic *hydrobiotite* are loose, soft, pale-green or, less often, greenish-yellow. They contain numerous inclusions of large (20-70 μm), idiomorphic, octahedral crystals of pyrite, which extremely strongly deform and break the mica into thin, crumpled plates. Inclusions of marcasite in hydrobiotite are occasionally found. Clastic *muscovite*, as a rule, is devoid of any inclusions. In the boundary layer with surface-oxidized rocks (10-15 cm thick), along with neogenic pyrite, relict corroded inclusions of crystalline goethite are preserved in clastic hydrobiotite, marking the former zone of surface oxidation.

Clastic *ilmentite* is devoid of any signs of leucoxenization; very rarely, fine octahedral pyrite is observed on the surface of grains. The portion of dark-gray and black *feldspars* in the rocks of layer 2 is 3-5 times less than in originally gray-colored ones, on account of appearances of grains of light-gray and white color.

Organic matter in the form of dull, earthy detritus often has an aureole of cinnamon-brown color in the surrounding rocks; phytomorphic iron disulfides are absent. According to results of thermal analysis, the carbonaceous detritus is classified in thermal group 1A, which characterizes oxidized organic matter [6].

Sandstones with basal *dolomitic* cement among whitish-gray rocks have a gray, dark-gray, or black color, often with a distinct brownish or reddish shade. The latter is connected with the presence in the middle part of rhombohedral dolomite crystals of residual pseudomorphic iron hydroxides with a rosy-orange color. In the outer parts of the crystals, fine (1-5 μm), platy aggregates of pyrite are concentrated, which are arranged by zones of the rhombohedra's growth. It is precisely these pyrite aggregates that give the dolomite and the sandstone as a whole a black color. In the pore space of sandstones with carbonate cement, large (30-100 μm) octahedral crystals of pyrite and marcasite are formed. When a large amount of iron disulfides appears, relict iron hydroxides are not preserved.

Thus, the main traits distinguishing the whitish-gray rocks of layer 2 from the originally gray-colored rocks of layer 1 are the complete absence of framboidal pyrite aggregates, the appearance of large, idiomorphic pyrite crystals in the intergranular space and in flakes of hydrobiotite, strongly deforming the latter, the oxidized state of carbonaceous detritus, the combination of relict forms of iron hydroxide and neogenic iron disulfides in rocks with dolomitic cement, and the combination of relict forms of crystalline goethite in hydrobiotite and neogenic pyrite in the intergranular space in the transitional zone to layer 3. On the basis of these mineralogical traits, the whitish-gray rocks can be characterized as secondarily reduced, having been formed as a result of superposition of epigenetic sulfidization on a zone of ancient surface oxidation.

The mineralogical criteria for diagnosis of such a zone can be used most efficiently when secondary processes of oxidation and reduction develop in sections where light-gray, diagenetically slightly reduced lithotypes occur, against the background of which secondarily reduced rocks do not stand out visually.

Variant 3. The structure and geochemical subdivision of the cross section of the lower rhythm is the same as in the preceding variant. Originally gray-colored rocks with the complete set of mineralogical traits of this genetic type contain an increased (up to 0.47%) amount of iron disulfides, as a consequence of the mass appearance of *marcasite*. The latter is represented by idiomorphic, solitary, short-prismatic crystals (5-30 μm) concentrated in the intergranular space of the rocks. The ratio of pyrite to marcasite is equal to 1:5-1:20. Sometimes, marcasite forms inclusions in hydrobiotite and muscovite; concretions of crystals and large (up to 200 μm), spherical, radial microconcretions with an aggregate of framboidal pyrite in the middle are found less often.

The increased content of marcasite in gray-colored rocks in amounts not characteristic of originally gray-colored formations enables us to classify them as preore-disulfidized. Disulfidization of originally gray-colored deposits occurred simultaneously with the reduction of rocks in the zone of ancient surface oxidation.

Variant 4. At the base of the rhythm (see Fig. 1c) lie originally gray-colored rocks (layer 1), which are replaced upward by oxidized orange- and reddish-yellow sands and sandstones with a mottled distribution of brighter-colored points of the same color. The upper part of the cross section is composed of secondarily reduced (layer 4) and surface-oxidized rocks (layer 5).

Microscopic study of sheet-oxidized rocks makes it possible to confidently divide them into two layers. The sandstones of layer 3 contain distinct pseudomorphs of iron hydroxides on large octahedral crystals of pyrite in the intergranular space and on inclusions of them in clastic biotite. The pseudomorphs possess ideal preservation. The greater part of the hydroxides forms a dense, cryptocrystalline *nucleus* of reddish-brown color, with very fine details (striation on the faces) duplicating the morphology of the replaced crystal of iron disulfide. Due to differences in the density of iron hydroxides and disulfides, approximately 5-10% of the hydroxide phase does not fit in the volume of the pseudomorphs. An insignificant part of it covers the surface of the pseudomorphs' nucleus that is free of contacts with clastic grains and argillaceous cement, in a thin (1-5 μm), loose layer, forming a *primary ochreous jacket*. The greater part of the released hydroxides settles in the form of an ochreous coating on the surface of nearby clastic grains or impregnates the argillaceous cement, forming a spherical, brightly colored, *ochreous halo* around pseudomorphs. The thickness of the layer of ochers near the nucleus is 10-30 μm ; toward the periphery it decreases. The number of colored grains in a radial direction reaches 5-10, and the size of the halos is from 0.1 to 1-2 mm. It should be emphasized that the macroscopically visible pseudomorphic nature of coloring of rocks in zones of sheet oxidation is connected with the presence of precisely such ochreous

halos: the nuclei themselves, which are true pseudomorphs in accordance with the mineralogical definition of this term, have a dark, reddish-brown color, insignificant dimensions, do not stand out in color among the other dark grains, and do not play the role of pigment.

In the central parts of rhombohedra of *dolomite*, rosy-orange, finely dispersed, iron hydroxides are preserved, which are left from changes in the mineral in the stage of surface oxidation. Fine, platy, pyrite aggregates located along zones of growth in the outer parts of crystals are pseudomorphically replaced by reddish-orange, dense hydroxides that polish well. Near the contacts with gray-colored rocks, relicts of incompletely oxidized platy pyrite are observed in the dolomite. There is apparently no decrease in the portion of idiomorphic bivalent iron in the composition of carbonate.

Flakes of *clastic biotite* are soft, plastic, light- and yellowish-green. Numerous inclusions of octahedral pyrite crystals contained in them are transformed into pseudomorphs, from which spots of finely dispersed iron hydroxides spread along cleavage planes, sometimes going beyond the bounds of the flakes and coloring grains adjacent to them. *Carbonaceous detritus* is black and dull. According to data from thermal analysis, it is classified in thermal group 3C, which characterizes oxidized and strongly transformed organic matter [6]. Clastic grains of *ilmenite* are leucoxenized to various degrees.

In the distribution of coloring of oxidized rocks, the following pattern is noted: a predominance of red shades is characteristic of relatively coarse-grained sands; yellow, of finer-grained and clay ones. These differences may be connected with the increased content of sporadic insets of rosy-red, oxidized Fe-dolomite in the former variety of sands and its practically complete absence in the latter.

These observations indicate that the rocks of layer 3 were formed as a result of the development of sheet oxidation on secondarily reduced rocks.

The oxidized sands of layer 2 are visually distinguished from overlying ones by a paler overall color and distinct structural confinement of pseudomorphic speckles concentrated mostly along thin interlayers with increased content of minerals of the heavy fraction, carbonaceous detritus, clastic micas, and argillaceous cement, and around clay balls. At the microscopic level, the differences become even clearer. Framboidal pyrite aggregates growing on clastic minerals of the heavy fraction and forming phytomorphs on carbonaceous plant detritus, fine-crystalline varieties of pyrite in the intergranular space, and flattened inclusions in clastic biotite are subjected to pseudomorphic replacement by iron hydroxides, as is marcasite. The pseudomorphs' preservation is ideal; therefore, the morphological types of the substituted disulfides are easily established. In Fe-dolomite, zonally arranged, finely dispersed iron hydroxides are observed. In all other respects, the nature of changes is analogous to the rocks of layer 3. Observations show that the sands of layer 2 were formed as a result of sheet oxidation of originally gray-colored sands or preore-disulfidized rocks. The difference between them can be established according to the absolute number of pseudomorphs on marcasite or the ratio of the latter with pseudomorphs on pyrite.

Variant 5. The geochemical structure of the lower rhythm's cross section is analogous to the preceding variant, however the sheet-oxidized rocks are visually distinguished by a vaguely mottled distribution of iron hydroxides and a microscopically different structure of pseudomorphs on disulfides in oxidized preore-disulfidized and secondarily reduced sands. The dense nuclei of pseudomorphs are partially granulated and transformed to ochers. The process begins from the surface of the nuclei: first they are smoothed out, losing their distinct outlines, vertices, and edges. Then, the dense nucleus is transformed into ochers to some depth. As a result, in the place of an idiomorphic pseudomorph there is a compact, isometric clump of ochers with unformed relicts of a dense aggregate of hydroxides in the central part. Granulation of the nuclei is accompanied by an increase in the amount of ochers in ochereous halos: the color of the latter does not change in this case. Pseudomorphs on pyrite enclosed in biotite preserve their primary form better than in the intergranular space.

Differences in the aggregate state of pseudomorphs for rocks of the same epigenetic type in variants 4 and 5 are explained by different ages of formation of the zones of sheet oxidation, which is confirmed by the history of geological development of the area of the deposit and adjacent regions.

Variant 6. In the cross section of lithologically homogeneous strata of the lower rhythm, gray-colored rocks are replaced by oxidized ones over a distance of 50 m along a line steeply cutting across the bedding and forming an elbow bench 15 m thick. According to a set of macro- and microtraits, the limonitized sands belong to the early zone of sheet oxidation, which developed on secondarily reduced rocks (see Fig. 1d).

Gray-colored deposits found at the same level have a distinct whitish shade and are outwardly reminiscent of whitish-gray, secondarily reduced sands. However, microscopic observations showed a somewhat different set of min-

eralogical traits of these rocks. They include a larger size of octahedral pyrite crystals, reaching 100-150 μm , the appearance, in a number of cases, of isometric marcasite crystals, the appearance in hydrobiotite of a large number of inclusions of octahedral pyrite crystals transforming the mica into unformed, fine-flaky, light-greenish-gray aggregates, the presence of leucoxenized minerals of the heavy fraction, and the presence of relict forms of iron hydroxides. The latter are represented by finely dispersed films on the surface of clastic grains within the bounds of calcite pisiform concretions. In the outer parts of the nodules and on their surface, fine-crystalline, neogenic disulfides are constantly present, often growing directly on the hydroxide films.

Thus, the whitish rocks can be interpreted as repyritized, having been formed as a result of sulfide reduction of rocks of the early zone of sheet oxidation; and the boundary between them, as an epigenetic one connected with a break in the sheet-oxidation process.

Variant 7. Sheet-oxidized rocks with a bright-orange color, with a pocketlike mottled distribution of hydroxide pseudomorphs (layer 2) are developed on whitish, repyritized sands (see Fig. 1e). Such a nature of the interrelation is established on the basis of complete correspondence of the morphology, dimensions, and distribution characteristics of pseudomorphs of iron hydroxides on disulfides in unoxidized, repyritized rocks (layers 1 and 3). Rocks of the zone of sheet oxidation formed on secondarily reduced and repyritized sands can be distinguished according to the greater number and size of pseudomorphs.

Variant 8. In the cross section of the lower rhythm, according to the diagnostic traits cited above, surface-oxidized (layer 6); gray-colored, secondarily reduced (layers 2 and 5); and sheet-oxidized (layer 3) rocks are distinguished (see Fig. 1f). The latter contact the gray-colored rocks through interlayers of brightened sandstones 0.5-1 m thick (layer 4; at the boundaries of layers 2 and 3 these rocks are not shown, due to their small thickness). The brightened sandstones are completely devoid of ocherous iron hydroxides: ocherous halos on the surface of terrigenous material and primary ocherous jackets on the nuclei of pseudomorphs disappear, and the argillaceous cement is decolorized. In the intergranular space and in flakes of biotite, the dense nuclei of pseudomorphs are preserved, but they are not discernable macroscopically. They often bear traces of intensive corrosion. Irregularly shaped depressions on the surface of the nuclei are not filled with ochers; fragments of the flat faces of crystals of former disulfides are sometimes preserved. In rare cases, fine-crystalline (5-10 μm) octahedral pyrite is formed in the intergranular space. The appearance of brightened rocks is connected with the process of epigenetic gley reduction of the frontal part of the zone of sheet oxidation.

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The methodological techniques used by the authors made it possible to establish the sequence of epigenetic transformations at the studied deposit, to determine their different directions, and to distinguish genetic types of gray-colored (originally gray-colored, preore-disulfidized, secondarily reduced, repyritized, and postore reduced) and oxidized rocks. It was also established that the formation of complex rhenium-selenium-uranium ore is connected with two stages of sheet-oxidation processes separated by a stage of epigenetic reduction. The uneven manifestation of secondary changes in the area of the deposit, in conjunction with primary lithological and facial peculiarities of aquiferous horizons, led to formation of infiltration ores with variable productivity and complex morphology. Prediction of such ore in various stages of geological-survey work can only be done with consideration of the whole set of epigenetic processes in the ore-bearing strata.

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MORPHOLOGICAL ANALYSIS OF TERRIGENIC QUARTZ IN LITHODYNAMIC AND PALEOGRAPHIC RECONSTRUCTIONS

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A survey of methods for morphometric analysis of psammite is given and, using one of these methods as an example, the possibility of finding sedimentation conditions and sources of fragmental quartz in recent sediment from different climatic zones and Neogene deposits of the northeastern portion of the Sakhalin Island is illustrated.

Morphometric analysis of fragments forming sediment and sedimentary rock is one of the oldest and most highly developed methods of lithological investigation. It has previously been perfected and applied principally for the purpose of studying coarse deposits. The roundness and shape of pebbles and their type of surface, when employed using other methods of environmental analysis (granulometric methods, measurement of pebble orientation), have served as the basic criteria for determining the transport pathways and deposition conditions of psammite. Methods have been developed for determining roundness and a large number of morphometric indices have been proposed on the basis of measurements of the three mutually perpendicular axes of fragments. Analysis of particles of the size of grains of sand has already been conducted by a technique developed for psammite using the same approach to estimation of the roundness and shape of grains, but the method has not entered into widespread use because of technical difficulties. Usually, only qualitative estimates of the roundness and shape of grains, together with two-dimensional schlieren measurements, are employed. In recent decades the number of investigations concerned with the study of the morphology of fine fragmental grains and the formation of such grains has sharply grown in connection with the development of electron microscopy and computer technology.

The methods of morphometric investigations of psammite that have been developed up to now may be divided into three groups: (1) two- or three-dimensional measurements of grain axes and of grain roundness in schlieren or in stereological images [15, 17, 19, 24, 26, 29]; (2) electron microscopic study of grain surface [18, 20, 21, 25, 27, 28]; (3) computerized separation of particles based on their behavior in a dynamically active medium [12, 14, 20, 23].

A basic condition for the application of morphometric analysis when used to solve particular scientific problems is the presence of characteristic features in the shape of the fragments and of their surfaces which the fragments have acquired in the course of the disintegration of parent rock or in transport under exogenic conditions.

TECHNIQUE OF INVESTIGATIONS

The purpose of the present article is to determine the capabilities of lithodynamic and paleogeographic reconstructions using as an example the analysis of a very narrow component of terrigenous matter — quartz measuring 0.250-0.315 mm in dimension. Quartz was selected for study because of its widespread occurrence in sediment and sedimentary rock, the relative ease with which it may be tested, its stability in chemically active media, and its physical strength, and density, which is close to the average for all sedimentary rock.

A fraction measuring 0.250-0.315 mm in dimension was selected on the basis of long-existing tradition in the field [14, 20], and also in view of aerohydrodynamic properties. Unlike grains of larger sizes (e.g., 0.4-0.5 mm), grains of this size virtually never turn into nodules when transported in water currents [3, 13, 22, 26], though, in fact, they do turn into nodules quite rapidly when borne by wind. Grading by shape and roundness is also observed to occur in air currents. In a water current only partial grading by shape occurs, with the flattest particles being isolated [22].

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TABLE 1. Mean Values of Morphometric Coefficients of Quartz in Fractions Measuring 0.250-0.315 mm from Regions with Different Sedimentation Conditions

Sampling site, facies, borehole interval, m	Sample number	Number of grains	$2c/(a + b)$	c/b	Roundness
Coast of North Africa					
Sahara Desert	1	125	0.720	0.833	0.42
Bay of Tunis	2	107	0.687	0.789	0.31
beaches of the Mediterranean Sea	3	105	0.691	0.795	0.43
Dunes from the coast of the Baltic Sea	4	108	0.680	0.783	0.37
Dune sand from the coast of northern Sakhalin	5	122	0.666	0.793	0.33
Granite weathering crust (Singapore)	6	304	0.671	0.780	0.00
Bottom sediments from the Red Sea	7	81	0.669	0.775	0.19
Bottom sediments from the shelf of South China Sea	8	2671	0.640	0.750	0.16
Bottom sediments from northeast Sakhalin shelf	9	570	0.640	0.727	0.17
Bottom sediments from northwest Sakhalin shelf					
Sakhalin:					
Gulf of Sakhalin	10	58	0.647	0.760	0.22
Tatar Strait	11	120	0.675	0.779	0.22
Delta front of Amur River (20 m) – Amur lagoon	12	109	0.616	0.723	0.12
Delta front of Mekong River					
depth 15 - 25 m	13	532	0.592	0.727	0.12
depth 5 m	14	100	0.544	0.634	0.06
Estuary of the Mekong River	15	102	0.646	0.755	0.07
Neogene deposits of the Northern Sakhalin shelf:					
0-470	16	150	0.672	0.767	0.21
470-1100	17	419	0.674	0.750	0.19
1100-1300	18	210	0.671	0.766	0.18
1300-1650	19	240	0.703	0.780	0.17
1650-2250	20	480	0.705	0.794	0.19
2200-3000	21	510	0.695	0.784	0.19

In the present study traditional morphometric analysis together with measurement of the long, intermediate, and short axes of the grains under a microscope using a mirror attachment [3, 29] is employed. Roundness is determined by comparison to a standard scale [11]. Several of the most commonly employed dimensionless morphometric indices are computed in order to describe the shape of the grains, including the Cailleux index of flatness and the index of elongation [16], the Jahnke shape factor, the Snead and Folk index of roundness [29], the Corey shape factor [17], and the two-dimensional index c/b [3].

A total of 30-50 grains were studied in each sample if the results were used solely to determine the statistical parameters; 200-300 grains were used if a more detailed analysis of the distribution was carried out. Calculations of the indices and of the statistical distribution parameters were conducted on a VT-20X minicomputer using programs written by O. V. Zaitsev [3].

Samples from different recent sedimentological environments were selected for the analysis (Table 1). Three of the samples (No. 1-3) were taken from the coast of Central Africa, from a site near the city of Tunis, which is also the terminus of sand transport by wind from the northern Sahara [30]. In the last two samples, in addition to desert sand that had drifted to the sea by different paths, there were also present populations that had entered due to erosion of certain types of ancient rock.

Sample 4 was taken from the near-pectinal portion of a dune situated along the eastern coast of the Baltic Sea, near the mouth of the Venta River, and sample 5 from wind-borne coastal and sea sand from a site north of the Sakhalin Sea in the region of the village of Moskal'vo. Sample 6 was taken from a granite weathering crust from the coast of the South Chinese Sea (Singapore). Sample 7 constitutes hydrothermal sediment from the basin of Atlantis II in the Red Sea in which quartz of the size we are studying is a foreign terrigenous (edaphogenic?) impurity and is present in insignificant

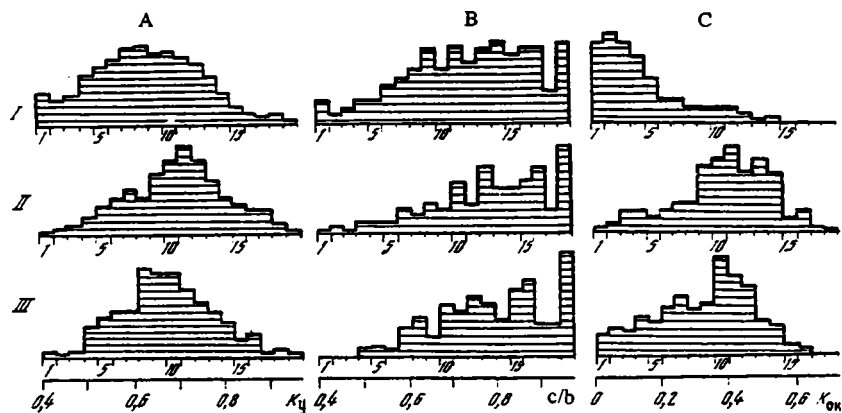


Fig. 1. Distribution histograms of the values of the morphometric index of flatness (A), the index c/b (B), and index of roundness (C) for Set I (combining 42 samples from the Sund shelf of the South China Sea), Set II (combining three samples from the coast of North Africa), and Set III (dune sands of the northern Sakhalin Island).

quantities. To describe the shelf deposits a total of 42 samples were taken from the Sund shelf of the South China Sea together with 19 samples from the shelf of the northeast Sakhalin Sea. They are represented by recent and relic late-Pleistocene sediment formed by drift from the Mekong and Amur Rivers, respectively, and material from local sources [2, 4, 8]. A total of seven samples of delta front deposits of the Mekong River were separately enriched (sample 13). Samples 10 and 11 constitute sections of the shelf contiguous to the Amur lagoon from the north (Gulf of Sakhalin) and from the south (Tatar strait). These sands, like those of sample 12 of the delta front deposits from the Amur lagoon, are composed of recent or ancient drifts from the Amur River, though aeolian processes are extensively developed along contiguous sections of the coastal region [8], and drift of sand that has undergone aeolian processes from the shelf undoubtedly occurs.

In order to describe the ancient conditions of sedimentation sludge from one of the oil-prospecting boreholes of the shelf of northeast Sakhalin that had become covered with deposits of a Neogene-epoch Nutovian level was studied. To simplify the representation of the results of the analyses, six of the samples are combined together in Table 1, with samples 16-18 characterizing a Pliocene-epoch Upper Nutovian sublevel and samples 19-21, a Miocene-epoch Lower Nutovian sublevel. Analysis of quartz from the sludge taken from this borehole did not involve measurements of split grains from fresh chips.

In our preliminary analysis of the significance of the morphometric indices we were employing and of their variability in recent deposits, it was established that indices that were based on a relationship between the three grain axes were identical [3]. Therefore, below we will use the Cailleux index of flatness $K_y = c/(a + b)$ and the index c/b , which reflects the cross-sectional shape of the grains (where a , b , and c are the long, intermediate, and short axes of the grain, respectively).

Generalized data on the shape and roundness of the grains are presented in Table 1. For a more detailed analysis the statistical moments of distribution are computed and histograms constructed in Fig. 1.

RESULTS OF INVESTIGATIONS

In most of the samples which we studied, the values of the indices of flatness and roundness exhibit an explicit trend towards a log-normal distribution (cf. Fig. 1). In some cases a polymodal distribution for the roundness is observed, associated with displacement of material from several sources. The values of the index of roundness in samples taken from different sedimentation environments correspond, to a first approximation, to the conclusions that had been reached by D. Goosens on the basis of experimental studies [22] concerning the predominant rounding of quartz of these dimensions when being transported by air currents. The optimal roundness is typical of samples that are completely or partially composed of material that has undergone aeolian processes, e.g., desert sediment (cf. Table 1, sample 1); dune sediment (samples 4

TABLE 2. Intervals of Histograms (cf. Fig. 1B) with Maximal Distribution Density of the Index c/b (underlined numbers refer to intervals with maximal distribution density for the particular sample)

Sample*	Number of peaks (numerator) and value of c/b (in denominator)					
	I	II	III	IV	V	VI
	0,49—0,55	0,58—0,67	0,67—0,73	0,73—0,82	0,85—0,90	0,94—1,00
Samples with highly rounded quartz						
1	4	7	11	13	17	19
2	3	5	9	<u>13</u>	17	<u>19</u>
3	4	7	11	<u>13</u>	17	<u>19</u>
4	4	7	10	<u>13</u>	16	19
Samples with poorly rounded quartz						
7	4	7	10	<u>13</u>	16	19
8	4	9	11	14	17	<u>19</u>
12	5	8	10	12	16	<u>19</u>
13	4	7	11	<u>14</u>	17	19
16	6	—	—	<u>12</u>	17	19
17	—	8	—	<u>12</u>	17	19
18	—	7	11	13	16	<u>19</u>
19	—	8	11	13	16	<u>19</u>
20	—	7	—	13	17	<u>19</u>
21	3	—	10	13	16	<u>19</u>
Samples with entirely unrounded quartz						
6	4	8	11	14	<u>16</u>	19

*For sampling site, see Table 1.

and 5), or coast-sea and shelf sediment taken from near shores with highly developed aeolian processes (samples 2, 3, 10, and 11). The minimal values of the index of roundness are exhibited by sediment from the delta of the Mekong River, which is a humid tropic zone (samples 13-15), where the rate of drift of material by means of water currents from recent weathering crusts is very high.

The values of the index of flatness in all the samples has a nearly normal or lognormal distribution that resembles a symmetric distribution. In terms of mean values, samples from aeolian and delta front conditions differ sharply, though these differences are not identical to the variation in roundness (cf. samples 14 and 15, Fig. 1). Obviously, differentiation between grains in terms of shape manifests itself in the course of transport. The flattest grains that are, correspondingly, also the most mobile in water currents are entrained by large rivers from dry land and deposited at sites where flow rates fall sharply as the rivers discharge into the sea. The most isometric grains are left in alluvium and even in estuarine deposits (sample 15). On the shelves material of the dimensions we are studying may travel and be redistributed in terms of shape only in rare instances and, usually, new morphometric sand populations are not formed, though old populations that arise with the entry of drifts from dry land are preserved. The exception is the case of ice dispersal in conjunction with wind-driven sand deposited on ice surfaces. In this type of transport it is possible for significant quantities of dune sand to drift for tens and hundreds of kilometers from the coast [1, 8]. In aeolian transport proper, isometric grains are very easily entrained, whereas flatter grains are retained within the ablation region, turning, mainly, into layers of mobile barkhan or dune sand.

The distribution of the mean values of c/b for samples from different sedimentation environments resembles the distribution of the index of flatness. The distribution within each sample, in contrast, proves to be more complex, exhibiting a number of peaks (populations) and a trend towards increased distribution density for values of c/b close to one (cf. Fig. 1B and Table 2). If it is recalled that the distribution of the values of c/b exhibit certain elements of lognormality, a feature that is, in part, apparent in the histograms, the mode of this distribution will be within the range 12-14 (cf. Fig.

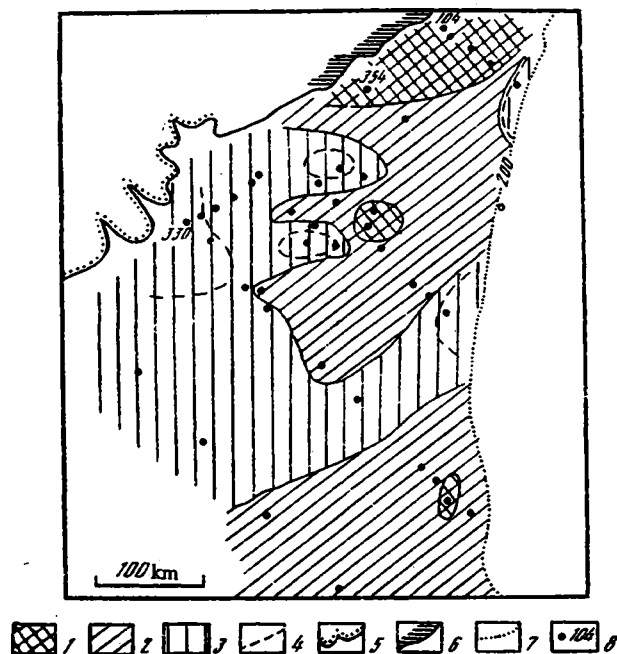


Fig. 2. Roundness of quartz in fractions measuring 0.250-0.315 mm from surface deposits of the shelf of the South China Sea contiguous to the mouth of the Mekong River: (1) >0.20 ; (2) $0.15-0.20$; (3) <0.15 ; (4) 0.125 roundness isoline; (5) delta of the Mekong River; (6) desertified sections of the coast in the Fanri region with highly developed wind-blown topographic features; (7) 200-m isobathic line; (8) samples of deposits studied and sample number assigned to each.

1c). In this case the peaks on the intervals 13-14 prove to be relatively slight and commensurate with other small peaks, for example, those on the intervals 4, 7, and 10-11. The peaks on the intervals 16-17 and 19, in contrast, are sharply distinguished against the background of the overall distribution, which is close to a lognormal distribution.

The presence of intervals with high distribution density c/b is attributable, in part, to the special crystallographic shapes assumed by quartz. The principal portion of the mass of terrigenous grains enters sediment from acidic magmatic and metamorphic rock in the form of primary xenomorphic grains. Moreover, they form, obviously, the principal population of grains with values of c/b in the range 0.70-0.85. In the analysis dipyrnidal or short-columnar crystals of β -quartz, often partially fused or rounded, that enter mainly from rich siliceous effusive rock, are often observed. Crystals of low-temperature α -quartz in the form of elongated pseudo-hexagonal prisms in combination with two trigonal bipyramids or fragments of the latter, rounded to varying degrees, are encountered significantly less frequently. Moreover, highly rounded or partially disintegrated grains that most probably originally were in the form of crystals are common. In the case of all the different types of crystals which we have been discussing, the cross section in which the values of c and b are measured is hexagonal in shape with calculated value of the ratio c/b of 0.866, which falls within interval 16 (0.85-0.88) in the histograms. In the case of rounded or partially fused crystals the value of c/b approaches one. Accordingly, it may be supposed that the peak V in the intervals 16-17 (0.85-0.91) in the c/b histograms is determined by the content of unrounded or slightly rounded quartz crystals, while peak VI (0.95-0.97), by partially fused or rounded crystals. This is confirmed also by the shift of the maximum of peak V to interval 16 in sample 6 (cf. Table 2), which is represented by entirely unrounded quartz from a weathering crust. Moreover, in all the samples with rounded quartz, peak VI is substantially higher than peak V.

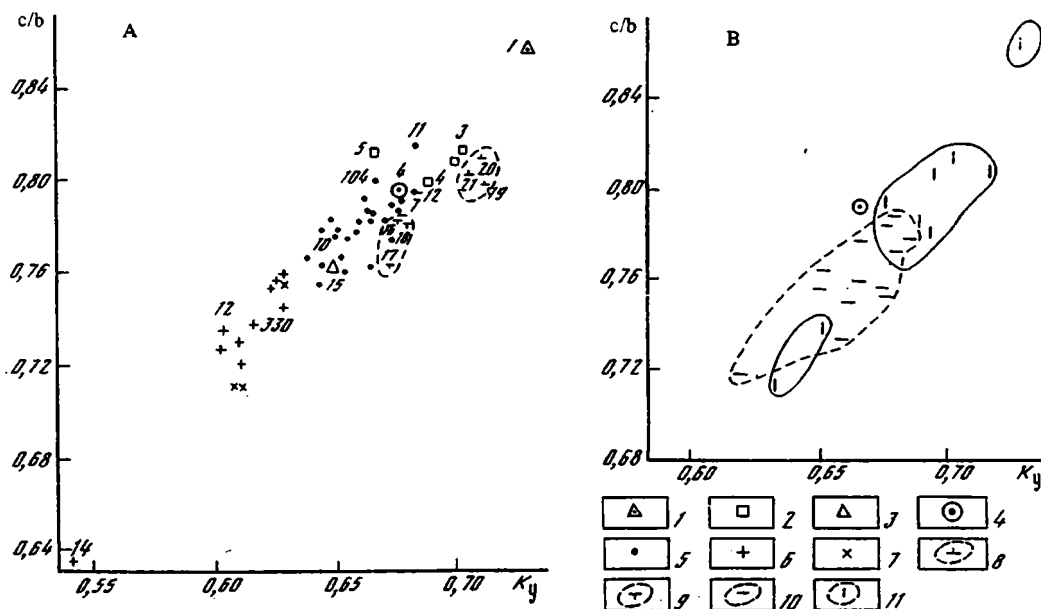


Fig. 3. Morphometric diagrams of quartz from recent sedimentation environments (A) and layers Ia and IIb from a Neogene-epoch Nutovian level found in the northeast portion of the Sakhalin Island (B): (1)-(7) samples of recent deposits (1, desert deposits; 2, dune and coastal and sea deposits from coastlines with highly developed aeolian processes; 3, alluvial deposits; 4, granite weathering crust; 5, shelf deposits; 6, delta front deposits; 7, relic delta front deposits of the Mekong River); (8)-(10) Neogene deposits from a borehole in the shelf of the northeast Sakhalin Island: 8, Lower-Nutovian sublevel; 9, Upper Nutovian sublevel; 10, Upper Nutovian sublevel, clayey band with aleurite intercalations (layer IIb); 11, Lower Nutovian sublevel, clayey-aleurite-sandy band (layer Ia). The digits in diagram A are the sample number (cf. Table 1 and Fig. 2).

SEDIMENTATION CONDITIONS AND MORPHOLOGY OF SEDIMENTARY GRAINS

An interpretation of the results of our morphometric analysis of terrigenous quartz may be directed towards either a search for the sources of supply of quartz or to establishing the sedimentation conditions. Usually, the two problems prove to be interrelated, and whether a well-grounded solution is possible for either of them depends upon the particular conditions of the region. Whether it is possible to establish the sources of terrigenous quartz depends upon the presence of the several morphometric populations of grains noted earlier, which are also partially related to the crystallographic varieties of these grains that have been preserved in the course of the rounding process (cf. Table 2). In view of the extremely low likelihood for quartz of the dimensions we are studying to become rounded in the course of transportation by water currents, roundness may be considered a diagnostic sign of grains from particular sources (under the assumption that transport is exclusively by water currents).

Characteristic variations of roundness of quartz grains from surface sediment determined by different sources of entry were discovered on the shelf of the South China Sea (Fig. 2). Here, the least index of roundness is typical of the region adjacent to the mouth of the Mekong River. In column 330 the sands were distinguished by homogeneity and a unimodal distribution of roundness with values that tended to zero. To the north, in column 354 and, especially, column 104, entirely unrounded quartz is lacking and stable populations with index of roundness in the range 0.08-0.16, 0.28-0.32, and others are found. The source of these populations may be sand that has undergone aeolian processes along the coast in the Fanri region. Here, because of the relatively arid microclimate, large sections of the land along the coast lack vegetation and both coastal and alluvial sands as well as material from weathering crusts are drawn into aeolian transport. Still another source of highly rounded quartz could be Cenozoic sedimentary rock that has washed out in coastal zones, on islands, and along ridges. The low value of the index of roundness of individual samples in the central portion

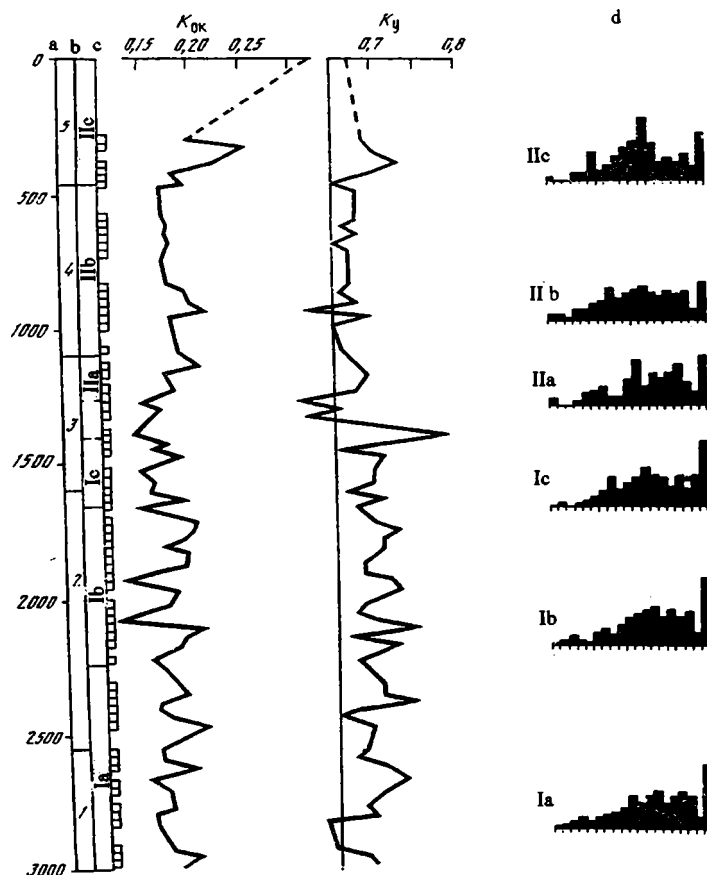


Fig. 4. Morphometric characterization of quartz in fractions measuring 0.250-0.315 mm by borehole from the shelf of the northeast portion of the Sakhalin Island: (a) litho-stratigraphic bands of deposits (1) and (2) Miocene-epoch Lower Nutovian sublevel: (1, aleurite-clayey band; 2, clayey-aleurite-sandy band); (3)-(5) Pleistocene-epoch Upper Nutovian sublevel: (3, aleurite-sandy band; 4, clayey band with aleurite intercalations; 5, sandy-aleurite band); (b) layers identified from the results of morphometric analysis of quartz in fractions measuring 0.250-0.315 mm; (c) testing intervals by sludge; (d) histograms of values of index c/b .

of the shelf and along the brow may be attributed to outcrops of relic Holocene and Late-Pleistocene delta front deposits of the Mekong River. The presence of these deposits here has been previously suggested on the basis of results of a lithological and environmental analysis employing other methods [4, 7].

The shape of the grains as described by the three-dimensional indices may also be determined from the origin of the quartz, though in most cases it depends on the degree of hydraulic (aerodynamic) differentiation occurring in the course of transport and sedimentation processes. This already is manifest upon comparison of the mean values of the indices for the aeolian and delta front deposits (cf. Table 1), though it is more clearly evident from the relation between the indices of flatness and the index c/b in the diagram (Fig. 3a). We were able to identify a rather compact field, the extreme points in which correspond to samples with quartz that is maximally homogeneous with respect to method by which the quartz is transported, e.g., wind-blown sand of the Sahara desert from the region of their ultimate destination (sample 1) and a series of samples of delta front deposits from the Mekong and Amur Rivers with quartz that has undergone lengthy transport and differentiation in water currents (samples 12-14). Points corresponding to samples with less pronounced degree of grading in one of the media or in a mixture of quartz that entered by various pathways are situated between these two extremes. Samples corresponding to dune deposits (samples 4 and 5) or coastal and sea deposits in regions with highly developed aeolian processes (samples 2 and 3) occupy the upper part of the field, and shelf deposits the lower part. Moreover, the field of shelf deposits is highly stretched out and, in principle, may be subdivided in greater detail on the basis of the ratio of material that has undergone transport in aeolian flows versus in water currents. Samples of shelf and

sea deposits in which aeolian material is present (samples 7, 11, and 104; cf. Fig. 3A) occupy the upper part of the shelf, contiguous to the field of coastal and sea deposits from regions exhibiting highly developed aeolian processes.

The point corresponding to a granitoid weathering crust is situated in the center of the field. If this is not a random coincidence, the distribution which we have obtained will illustrate the different trends in shape grading or the capacity of grains of different shapes to undergo transport processes in water currents and air flows. This may be explained if it is assumed that grains of the dimensions we are studying are transported in water currents in suspensions, and in air flows by rolling and saltation.

MORPHOMETRIC ANALYSIS OF NEOGENE DEPOSITS FROM NORTHEAST PART OF THE SAKHALIN ISLAND

Points corresponding to combined samples from individual lithological bands of Neogene – epoch Upper and Lower Nutovian sublevels from the northeastern part of the Sakhalin Island are situated in the diagram (cf. Fig. 3A) to the right of the fields of points for recent deposits. One possible cause could be our use of sludge in which some of the grains split in the course of drilling. Since the flattest and most elongated grains that had split along the a axis are the least stable under mechanical action, grains with maximal values of K_v were eliminated from the analysis. This is most likely the reason for the partial absence of peaks I, II, and III with minimal values of c/b from the histograms for c/b (cf. Table 2) in the case of the combined samples 13-18. If splitting of grains is taken into account, the fields for the Neogene deposits (cf. Fig. 3A) will, in fact, be situated to the left and may be correlated with fields of recent shelf and coast and sea deposits.

In the initial analysis of the morphometric indices several layers were identified in each borehole on the basis of their values or on the basis of the nature of the variations (Fig. 4). The greatest differences were observed between the groups of layers I and II, which, to a first approximation, are comparable with the Upper and Lower Nutovian sublevels. They are reflected in the mean values of the indices and are clearly visible in each section (cf. Table 1, Fig. 4).

There is a sharp variation of the morphology of quartz in the interval 1300-1400 m. It is clearly traced in the variation of the values of K_v and also from the histograms for the index c/b (cf. Fig. 4). In the latter case layers I are characterized by the predominance of the population (peak) VI (interval 19-20) and a nearly lognormal distribution in the remaining part. In layers II population (peak) V (samples 11-12) predominates and additional maxima are found (samples 5, 7, 16-17). Based on the hypothesis that the Amur River was the source of the sandy material used in the formation of the Neogene deposits [8, 10], two possible causes for the differences in the morphology of quartz may be identified. Either there were substantial variations in the watershed (climatic, geographical) early in the Pliocene and the Amur River began to deposit material of a different composition, or the environmental conditions in the sedimentation basin changed radically. The latter hypothesis is based on existing interpretations of sedimentation conditions in the Neogene epoch [6, 9, 10], according to which a Lower Nutovian sublevel was formed on the continental slope, while an Upper Nutovian sublevel was created within the shelf. Our data correspond more to this model, though in order to interpret the data it is necessary to make use of additional data on recent sedimentation in the Sea of Okhotsk.

Interpreting the results for group II of the layers does not entail any difficulty. In the diagram (Fig. 3A) the combined samples of these layers, with allowance for correction (cf. above), falls within the field of the shelf deposits. Points for the individual samples of layer IIb (cf. Fig. 3B) also coincide with the field of shelf deposits, while some of the samples fall within the delta front field. Points of layer IIc are shifted somewhat in the direction of the field of coastal and sea deposits. In view of the greater rounding and the variability between sections, it may be supposed that this layer was formed in a shallower portion of the shelf, at the point where material from aeolian landscapes entered during regressive periods.

The combined samples of layer I proved to be close, in terms of morphometric features, to coastal and sea deposits containing material that had undergone aeolian processes. Analysis of individual samples (cf. Fig. 3B) and of the variability between sections (cf. Fig. 4), however, reveals an extraordinarily high scatter with respect to both the roundness of the grains and with respect to their shape. Most of the samples from layer Ia, for example, form a compact field that coincides (with some correction) with the coastal and sea deposits, though there are individual points that gravitate towards both the delta front and the typical aeolian deposits. In view of the formation of deposits on continental slopes, this may be explained by the rhythmicity of the entry of terrigenous material during regressive and transgressive periods. In the course

of retreats, the shore line and mouth of the Amur River would have approached the brow of the shelf and sandy material from the delta front and coastal dry land could have drifted to the continental shelf by wind and in suspension flows. From an analysis of recent sedimentation conditions [1, 2, 8], in transgressive periods, ice dispersal would have been the basic source for the supply of sandy material within the area of the continental shelf. Wind-driven sands from the northern Sakhalin Island that drifted to the surface of fast ice [1, 8] would have been the most actively involved.

Thus, the greater part of the variability of the morphometric indices by section of the Lower Nutovian sublevel may be attributed to entry of sand from different environmental zones (aeolian, coastal and sea, delta front) into the continental slope during regression and transgressive periods.

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The investigations which we have conducted show that morphometric analysis of terrigenous quartz can be a rather successful means of solving a number of paleogeographic and lithodynamic problems in conjunction with other methods. The presence of several crystallographic forms of quartz in primary material that enters the exosphere, differences in the rate of rounding of grains and grading by shape in the course of transport by means of air flow and water currents are the foundation for this approach in investigations of fractions measuring 0.250-0.315 mm. Based on these factors deposits are identified that were formed under different dynamic conditions, a result that, in a number of cases, may also serve as a basis for identifying the individual facies in which these conditions predominate. Obviously, it is necessary to take into account the results of other methods of environmental analysis.

The use of morphometric analysis for subdividing sand deposits that were formed under the coastal and sea conditions of the tropic and subtropic zones is the most promising approach. In detailed areal testing and investigations of Quaternary deposits it has been possible to discover the sources, means, and pathways by which sandy material is transported to a shelf and, in certain cases, to make a quantitative estimate of these processes. The delta deposits of large rivers, eroded magmatic, metamorphic, and sedimentary rock and recent and ancient aeolian formations are among the potential sources that contain characteristic forms of quartz.

In the analysis of the morphometric features of quartz, it is possible to make a lithostratigraphic correlation between thick sedimentary strata and both large cycles that have arisen due to tectonic reconstruction as well as small cycles related to eustatic variations of the sea level. These structures are possible also in the analysis of sludge from oil-prospecting boreholes, though in correlating the results to models of recent deposits it is necessary to adjust the values of K_y , taking into account cleavage of some of the flattest grains. For the borehole we have studied, which is situated on the shelf of the northeast portion of the Sakhalin Island, an area that is covered by slightly condensed Neogene deposits, the correction factor is in the range 0.93-0.97.

The capabilities of morphometric analysis may be extended by studying quartz of several different dimensions. For diameters less than 0.10-0.15 mm, quartz virtually never becomes round nor graded by shape, i.e., its morphometry is determined solely by the primary composition. Large fractions (greater than 0.4-0.5 mm) become round and may become differentiated in water currents, and their morphometry is determined nearly exclusively by the distance and length of the transport process.

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